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Editor-in-Chief
Boris S. Sokolov

Member of the Russian Academy of Sciences,
Counsellor of the Presidium of the Russian Academy of Sciences (RAS)
Address for correspondence:
Staromonetnyi per. 22, Moscow, 109180 Russia
Phones: 7 (095) 951-2164; Fax: 7 (095) 953-5590.

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vtopor@paleo.ru

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aroza@paleo.ru

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P.O. BOX 1831, BIRMINGHAM, AL 35201-1831 USA
PHONE: (800) 633-4931; (205) 995-1567 (outside US and Canada)
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COMPANY "NAUKA/INTERPERIODICA"
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Stratigraphy as Geochronometry

A. V. Popov

St. Petersburg State University, Shestnadtsataya liniya 29, Vasil'evskii Ostrov, St. Petersburg, 199178 Russia

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Abstract—The present development of stratigraphic methods, especially of zonal stratigraphy, necessitates usage of a correct measurement procedure in stratigraphic investigations. Such a procedure should be based on the independent external metering system, e.g., on the Golovkinskii principle. An important component of stratigraphic investigations, all efforts are ultimately aimed at, is the establishment of boundary stratotypes. The main and single goal of stratigraphy is to elucidate the geological time structure. The most complete reconstruction of geological processes in time and space is achieved by geological mapping that represents, together with stratigraphic investigations, an integral method of graphic modeling of geological development in a form of geological map.

Key words: structure of stratigraphic investigations, principles of stratigraphy, boundary stratotypes, chronostratigraphy, ecostratigraphy, classification structure.

INTRODUCTION

The last decades were characterized by increasing attention to theoretical aspects of stratigraphy. The individual stratigraphic methods, geological and biological phenomena, the methods rest on, as well as stratigraphy, as a logically organized system of methods, were comprehensively analyzed. The intensive theoretical studies were stimulated by significant achievements of zonal stratigraphy and by the necessity to develop the global stratigraphic standard, i.e., the Standard Scale (SS). The growing interest to stratigraphic fundamentals is evident from elaborated new concepts and from different approaches to stratigraphic problems. In the 1970s–1980s, there was a distinct tendency to study the structure of stratigraphy, to consider a combination of its interrelated principles and methods, which are aimed to date geological events.

Works by Sokolov (1974, 1986, 1991) and Meyen (1981) who defined principles of stratigraphic procedure were of great importance for elucidating the system of stratigraphic investigations. In spite of some similarities, schemes of stratigraphic procedure proposed by Sokolov and Meyen were essentially different. Zhamoida (1980, 1984) also worked on the same problems. Ideas of Stepanov (Stepanov, 1954; Stepanov and Mesezhnikov, 1979) were principal for development of theoretical stratigraphy. He proved the necessity of chronostratigraphic units and considered biostratigraphic method as principal one for external metering.

The greatly increased resolution and precision of zonal stratigraphy, a wide arsenal of organic groups studied, and new effective stratigraphic methods allow us to use more correct principles of investigations. Important achievements of stratigraphic methods open the possibility to employ the correct procedure of measurements in stratigraphy.

PRINCIPLES OF GEOCHRONOMETRY

It is necessary first to touch upon the main rules of metering in science, which are often neglected in stratigraphic practice. We should distinguish between the notions of comparison and measurement. Any object can be understood through its difference from and similarity with related objects, and through elucidation of its internal differentiation. A correct comparison should meet the following requirements: first, objects and phenomena under comparison should be of the same kind and, second, only essential features of the objects should be compared.

A simplest comparison means an immediate correlation of phenomena and objects by means of some features. The measurement, as a perfect way of comparison, implies that phenomena and objects can be correlated to each other through a system of standards, i.e., of quantitative metering units of a necessary quality. The measurement enables a more accurate and comprehensive correlation of phenomena and objects. Like any other correct study, the stratigraphic investigation should follow these general rules.

Investigations (measurements) are important in man's activity. Like the investigation itself, a measuring tool stems from a conventional idealized scale though based on real objects and phenomena. Like any measuring procedure, stratigraphic investigation cannot go without a metering tool, and this tool is of basic importance for stratigraphy. A general approach to strati-

graphic investigations depends on the meaning we read into this notion.

Stratigraphic investigations employ two essentially different measurement methods based on different metering tools. One of the methods (endochronometry) is based on the internal heterogeneity of the object under investigation (for example, lithostratigraphy). The other one (exochronometry) uses the external metering system such as the organic evolution or radioactive decay (Popov, 1973, 1993).

Qualitative non-uniformity of geological events is a privilege of endochronometry, whereas their quantitative characteristics are investigation objects of exochronometry. This distinction is very conventional but necessary for the purpose of analysis. The time–space development of the geological process can be revealed only through the integrated endochronometric and exochronometric studies in combination with geological mapping, if intend to elaborate a dynamic time–space model for geological development of matter. In essence, the harmonic integrity of stratigraphic investigation and geological mapping represents a method of time–space modeling of the geological process in a form of resulting geological map.

Regretfully, the role of W. Smith, the author of a complex time–space modeling of the geological process, is underestimated. Any geological map showing age subdivisions of rock strata reflects general peculiarities of the geological process in space and with time. The interconnection of time and space, that is a fundamental feature of matter, is brightly expressed in a geological map.

The present-day stratigraphy concentrates efforts on the qualitative aspects of the geological process, whereas the quantitative ones do not receive the adequate attention, because some stratigraphers negate the absolute time notion. This eliminates in principle a possibility of using the correct measurement procedure in stratigraphy.

A correct measurement procedure (including that of stratigraphy) includes certain demands to the metering tool. A process chosen for the exochronometric tool must be autonomous in space and isochronous in time. It must be independent of an object under study. To eliminate (or to diminish significantly) the tool–object interdependence is one of the most essential prerequisites to obtain a pure result.

These characteristics of an ideal metering tool correspond to the Newtonian time. None of the natural process with these characteristics is known to science. Therefore, stratigraphers have to invent such a tool approximating the ideal one. The characteristics indicated above are inherent in the organic evolution and other geochronological instruments. Due to autonomy of the evolutionary process, we have a tool of stratigraphic investigations, which, however, is not a mold of the evolution proper.

The correct measurement of geological time inevitably demands introduction of such notions as the ideal (absolute) time and ideal (absolute) stratigraphic tool. They should be supplemented by the notion of absolute synchronism, approaching to which is the main goal of stratigraphy by geological mapping.

The Standard Stratigraphic Scale based on biostratigraphic records functions as the absolute (ideal) metering tool in the Phanerozoic stratigraphy. The organic evolution has such a fundamental feature as the considerable autonomy of the process. At the same time, some peculiarities of the organic evolution allow elaboration of detailed and interconnected zonal scales representing a kind of an isochronous field. Although the autonomy of the organic evolution is not self-evident in opinion of some researchers, it has been used since the time of early stratigraphic investigations as the main tool of exochronometry.

Leaving aside the problem of autonomy of the organic evolution (Popov, 1973, 1989), I should note that simple physical factors, such as changes in temperature and salinity of water and others, could not cause the development of one-celled organisms into the highly organized ones, including a man with his mind. This disproportion between the magnificent results of organic evolution and primitive effects of abiotic factors contradicts the principle of adequate response of living organisms to environmental influence.

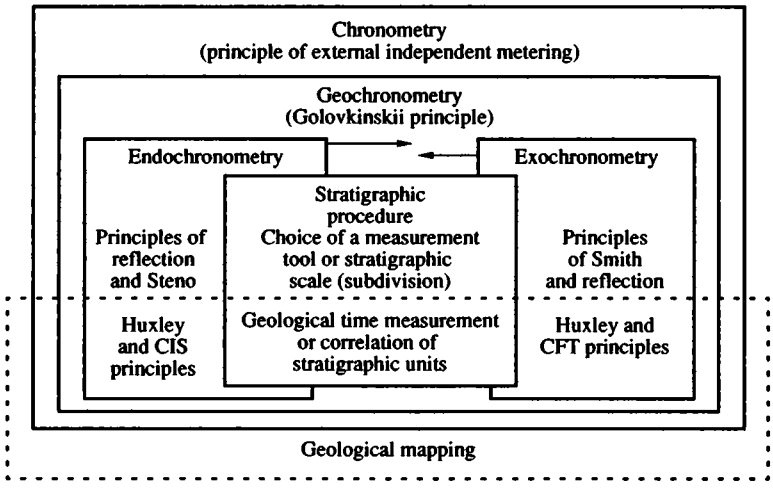
It should be emphasized that the stratigraphic scale cannot and must not reflect the time–space structure of biological and, especially, geological processes. It is aimed to be a tool that depicts the time–space peculiarities of two specific (geologic and biologic) forms of matter evolution. The same statement is correct in general with respect to stratigraphic schemes and individual stratigraphic units, including the lithostratigraphic ones. They are closely related to biological and geological events, as they are based on some manifestations of these events. However, they are not the adequate reflections of the latter.

Stratigraphic investigations include two operations. One of them is a subdivision of a rock or faunal (floral) succession into sequential units using the chosen metering tool. This operation includes the identification and dating of fossils. The other operation is correlation, i.e., the metering procedure based on the established subdivision scheme that is taken as a standard.

These stratigraphic operations are based on a set of stratigraphic principles. Different researchers accept different principles. Let us consider those principles, which seem to be necessary and sufficient for stratigraphic investigations. A combination of stratigraphic principles should fit the logically consistent procedure. For instance, the presented graphic scheme (Table 1) illustrates principal interrelations between main components (principles) of stratigraphic procedure.

As stated above, chronometry that is based on the external metering system grounds the correct normal

Table 1. Structure of stratigraphic investigations in connection with geological mapping



measurements. In geology, this is geochronometry based on the Golovkinskii principle, which should be considered in detail.

Golovkinskii (1868, p. 399) formulated that "...one should keep in mind a distinction between chronological, stratigraphic, petrographic and paleontological horizons." Further, he noted that "...chronological horizons cross indirectly all others..." (Golovkinskii, 1868, p. 400). These formulations imply that Golovkinskii not only realized the intricate time relationships between the process of the geological and organic evolution, but he also properly understood the goals of stratigraphy. In his terms, the interrelation between geological events is expressed by means of their time characteristics, which are most fundamental for the events.

Later on, the Golovkinskii principle received a new "modern" interpretation simply implying "diachronism of lithostratigraphic units." This interpretation essentially reduced the content of the original principle. It does not include stratigraphic and paleontological horizons, and thus simplifies a complex nature of geological time and methods of its study. Nevertheless, the Golovkinskii principle is still of great importance for general geology even in this interpretation, although many researchers (Sadykov, 1974; and others) do not refer it to the main principles of stratigraphy.

The complex differentiated character of the geological time and corresponding ways of its study, which logically followed from the brief definition of Golovkinskii, should be developed further. Golovkinskii indicated the main components and their interrelations, based on which we should form the whole system of stratigraphic investigations.

The most important consequence of the Golovkinskii principle was the necessity and possibility of using in stratigraphy the normal correct measurement procedure based on the external metering system. It was the organic evolution, which can serve as such a system. It should be noted that the paleontological method was used in stratigraphy before and after the work of Golovkinskii, but I speak about the elaboration of a consistent, internally logical system of stratigraphic investigations with most successful usage of the unique events in the organic evolution as the external metering tool.

Merits of the Golovkinskii principle become more obvious when it is compared with the second basic principle of stratigraphy, i.e., with the Steno principle, which is basic one in lithostratigraphy in terms of the internal metering system (endochronometry). The Steno principle states that each bed is older than the bed above it in any sequence of sedimentary strata that has not been overturned. This principle transforms spatial relations into temporal ones. Division of a sequence into a number of successive stratigraphic units is the operation or tool of stratigraphic (time) measurements. In this case, the accepted subdivision scheme becomes alienated from the deposits, to which it is applied.

However, the Steno principle based on heterogeneity of deposits, i.e., on non-uniformity of the geological process itself, is of a limited accuracy that is characteristic of endochronometry in general. This defect is evident owing to the Golovkinskii principle indicating the facies diachronism and, hence, the limited abilities of lithostratigraphy. Using the Steno principle, we can precisely date a continuous section only. In this connection, Golovkinskii (1868, p. 397) wrote: "It seems rightful to put forward a seemingly paradoxical thesis: *common conviction in successive formation of beds successively overlying each other is erroneous.*"

The limited applicability of the Steno principle, which was shown by Golovkinskii, has not been recognized by all stratigraphers (Meyen, 1981). For example, Sadykov (1974, p. 56) considered that his thesis is a refutation of the first law of stratigraphy and nothing else.

Concerning the Steno principle, Golovkinskii (1868, p. 397) wrote "...this view is correct only for very limited areas...." Golovkinskii denied the absolute character of the Steno principle and indicated the narrow field of its usage. He also suggested some limitations for paleontological horizons. This is of principal importance for understanding the problem of geological time.

The Golovkinskii principle is of fundamental significance for stratigraphy. It is very important that Golovkinskii not only demonstrated the differentiated character of geological time, but also approached the necessity of absolute time notion for the correct stratigraphic investigation.

Another important consequence of the Golovkinskii principle that indicates diachronism of lithological horizons introduced a spatial aspect into the time characteristics of the geological process. It was a condition necessary to perceive interconnection between time and space. The Golovkinskii's idea was most thoroughly understood by Vernadsky (1975), who was first to state in 1885 on the unity of space and time.

Thus, the Golovkinskii principle does not rejected the Steno law but includes it as a constituent providing initial data on the lithostratigraphy and organic evolution. Afterward, these data are used in essentially different system of stratigraphic exochronometry, which cannot be reduced to the mechanism of the Steno principle.

Stratigraphic investigations based on the Golovkinskii principle follow the rules of correct measuring, which enable a deeper insight into the time and space characteristics of the geological process. In opposite, stratigraphic investigations based on the Steno principle only operate with an approximate undifferentiated notion of geological time possessing the one-vector characteristic. The application of the Golovkinskii principle suggests a notion of complex many-vector system of time and space for the geological development of matter.

The transition from the Steno principle to the Golovkinskii principle in stratigraphic investigations marks a turning point between two successive stages of stratigraphy development. From the methodical standpoint, it means a transition from direct correlations to measurements based on a system of time standards. Ideas of Golovkinskii were most fully developed by Stepanov (1954) who was first to establish chronostratigraphic units corresponding to certain time intervals and independent of physical properties of deposits. Later on, Stepanov and Mesezhnikov (1979, p. 264) concluded that the organic evolution could serve as the external metering system, and that it is methodically doubtful to measure events by means of the events themselves.

The principles of Golovkinskii and Steno are basic for stratigraphic investigations. They are combined with subordinate principles into the procedure of stratigraphic measurements. Table 1 shows that modern stratigraphic investigations are based on the correct metering (geochronometry) introduced into stratigraphy by the Golovkinskii principle. Stratigraphic investigation consists in two interacting approaches (endochronometry and exochronometry), which use essentially different procedures and tools of metering. Endochronometry plays an important though subordinate role allowing stratigraphy to normally function in the form of exochronometric system.

Basic in the endochronometry is the reflection principle by Sokolov (1991, p. 8) who stated the following: "... a stratigraphic unit that has stratotype and can be correlated and mapped within some ancient sedimentary basin is a reflection of the *real* geological setting and its natural limitations within the basin." In terms of geochronometry, the reflection principle is most effective for lithostratigraphy and, especially, for the Precambrian stratigraphy. In terms of exochronometry, the reflection principle is effective only for recognition of zonal units representing the elementary basic subdivisions. These units create the physical basis of the Standard Scale (SS) and stratigraphic schemes, which are instruments of exochronometric investigations.

The introduction of the reflection principle into stratigraphy necessitates admitting of a dual (conventional and material) character of stratigraphic units. Many of researchers who support now the concept of natural stratigraphy admit that the SS units and their boundaries are conventional to a great extent. However, all other units are subjects of discussion. The essence of dualism principle means diminishing controversies (differences) between the material and the ideal, or more exactly, a tendency to approach the ideal. This problem, which is fundamental for stratigraphy, is considered in detail below.

In exochronometry, operations of subdivision and correlation (Table 1), i.e., the elaboration of a metering instrument, is based on the Smith principle suggesting the independence of organic evolution. The modern interpretation of the principle is as follows: "Deposits may be differentiated and correlated by means of their fossil content" (Stepanov and Mesezhnikov, 1979, p. 59). The specific feature of organic evolution implies its independence of the geological processes and irregularity and unidirectional development of phylogenetic changes. Subdividing a paleontological succession we create a scale of biostratigraphic and corresponding geochronostratigraphic units. Radiological chronology is the main method of the pre-Vendian stratigraphy, which is outside the application sphere of biostratigraphic method.

The second operation of stratigraphic procedure is spatial tracing of time boundaries separating stratigraphic units, using the standards (stratigraphic units) obtained after the first operation. In endo- and exochronometry, the tracing of stratigraphic boundaries is based on the principles of homotaxy and correlative function transmission (CFT). These principles enable geological mapping, results of which reveal a complete

time–space structure of the geological development. These principles complete a set of the main principles of stratigraphy, among which the Golovkinskii principle is of organizing significance.

The homotaxy principle is uniformly understood as facilitating the rock dating by means of correlation of identical fossil successions, but the CFT principle needs to be considered in detail. There are two seemingly similar interpretations of this principle (Sokolov, 1986; Meyen, 1981; Zhamoida, 1984). The following interpretation of Meyen (Meyen 1981; Zhamoida, 1984) has gained a wider recognition: "Different, partly overlapping spatial complexes of stratigraphic indications imply their chronological interchangeability that can serve a basis for the intra- and interregional, up to global, correlations by means of the most important signs." This interpretation is known as the principle of chronological interchangeability of signs (CIS). It corresponds to the time correlation in terms of the "natural" stratigraphy, which considers deposits as having the same age, if they are related to the same event that necessarily happens within some time interval.

This interpretation does not suite the accuracy of correct stratigraphic measurements based on the external metering system, in which the ideal synchronism (boundary) means no time interval at all (Popov, 1993). The indicated system is more compatible with interpretation of Sokolov (1974, 1986) who suggested the principle of correlative function transmission (CFT) in connection with determination of the Silurian–Devonian boundary by "a golden spike." The CFT principle means, from the standpoint of measurements, a transmission of age information from a standard to an object thus metering the latter. To the contrary, the CIS principle suggests only an exchange of age information between objects thus determining a time span of a stratigraphic unit that characterizes a geological (biological) event. Such interpretation adequately reflects a spirit of the natural stratigraphy, which actually denies the measuring procedure.

However, practice of zonal stratigraphy inevitably implies necessity to use the metering procedure in stratigraphic investigations, as it is clearly evidenced by the international efforts to establish the boundary between the Silurian and Devonian systems. The ideas of "golden spikes" and the CFT principle are difficult to overestimate: they "opened new horizons for stratigraphy" (Sokolov, 1991, p. 9). The application of the "golden spike" and CFT principles was in essence an introduction of metering elements into stratigraphy. This became especially obvious when boundary stratotypes were established outside stratotype areas of the corresponding units. Sokolov (1986, p. 13) noted that the choice of the standard Silurian–Devonian boundary outside the stratotype areas of these systems, and even of their stages, had an astonishing effect. This decision, meaning a transition from fixation of time spans characterizing units to establishment of boundary stratotypes, was of great significance. First, stratigraphic investigations became considerably more precise; second, the separation of a boundary stratotype from a unit stratotype implies a certain autonomy of boundaries and their importance for calibration of stratigraphic scale; and, third, the choice of boundary according to agreement has been legalized. All these peculiarities are in good correspondence with a correct metering procedure.

The introduction of metering procedure into stratigraphic investigations posed the problem of boundary set. Sokolov (1974) noted that in spite of unique biological content of stratigraphic units, we cannot define accurately and unambiguously the boundaries between stages, systems, and erathems, and the task should be restricted. Sokolov (1974, p. 12) concluded: "there could be many boundaries in reality but we should choose a single one for the standard scale to facilitate in a greatest extent the mechanism of stratigraphic correlation."

The ideas of "golden spike" and the CFT principle are well correspondent to the notion of ideal synchronism that is principal in the correct metering system. Ultimately all practical and theoretical efforts in the field of stratigraphy are aimed to establish the synchronism. Achievements in this sphere influence the whole construction of stratigraphy and represent natural development of the correct metering procedure. In modern stratigraphy, this tendency is seen in transition from the unit to boundary stratotypes, in other words, from standardization of subdivisions (time spans) to fixation of standard boundaries. Significance of this transition, which took place when the Silurian–Devonian boundary was defined, is difficult to overestimate. For the first time, results of intense practical investigations have been theoretically generalized almost at once in the concepts of "golden spike" and CFT. This event was of great significance especially for practical stratigraphy. The priority of boundary stratotypes and the related theoretical concepts mark an important step of stratigraphy toward chronostratigraphy.

In terms of natural stratigraphy, the "golden spike" principle led, however, to some negative consequences. The lowered significance of unit stratotypes allowed them to be easily replaced, frequently without necessary substantiation, in the Standard Scale or stratigraphic schemes, in which they indicate particular time ranges. This noticeably undermines the most important feature of the SS, i.e., its stability. It is declaration of conventional character of stratigraphic units that can prevent unsubstantiated changes of traditional unit names and lead to normal detailing of the SS and stratigraphic schemes.

Thus, the above definitions of the CFT and CIS principles show their distinctions. Accordingly, the CFT principle can be incorporated in the structure of geochronometry, whereas the CIS principle seems to func-

tion in the sphere of endochronometry, which is directly based on geological events.

The necessity to clearly discriminate these principles is dictated by intense development of theoretical stratigraphy, the basis of stratigraphic correlation in which is still a subject of keen debates. Certain attempts have been made to develop the up-to-date theoretical base of the natural stratigraphy. One example is work by Simakov (1997) remarkable owing to a large scope of theoretical materials presented in a very complicated way and containing modern interpretations of well-known principles of the natural stratigraphy.

Let us consider basic theses of this work, leaving aside the whole massif of wordy considerations. Simakov (1997, p. 149) wrote: "Chronostratigraphic units correspond to sedimentary sequences deposited in the periods of a relatively stable (mobile–equilibrium) state of a paleosystem, the evolution of which is chosen as a basis for the scale elaboration." Further on: "Chronostratigraphic boundaries record events changing quasistable states of the paleosystem, the discrete-continuous development of which is a basis of metering" (Simakov, 1997, p. 149). It is obvious from both statements that Simakov follows the main course of the natural stratigraphy that considers geological events as ready metrics of a stratigraphic unit or a stratigraphic boundary. As usually, the stratigraphic unit meters a geological time span (a stable period of paleosystem) and the boundary records an event denoting a change in the paleosystem. The following definition of Simakov is especially demonstrative: "Golden spikes" are nomenclature standards (symbols) of natural chronostratigraphic boundaries and approximate chronological ecotones to the customary and suitable form of "dimensionless" points, which separate adjacent units on the conceptual time arrow." This definition brings down the ideas of "golden spike" and CFT principle, which have essentially modified the stratigraphic investigation procedure, to the *customary* and suitable dimensionless points separating stratigraphic units.

Accordingly, the important moments of development of stratigraphic methods, as products of specific man's activity, appeared to be out of the Simakov's attention. As a result, the newly elaborated effective tools of measuring (zonal scales and the CFT principle) have to maintain the considerably less precise metering system.

As before, basic in his system is the following thesis: "the systematic organization of the Nature itself dictates us an objective necessity to bring our theoretical (conceptual) considerations in correspondence with observable phenomena of reality, making some corrections by means of non-trivial conventions" (Simakov, 1997, p. 155). He believes that "general principle" of stratigraphy and relevant "operational apparatus" of stratigraphic measurements are to be reduced to a simple copying of secular variations in the evolving paleosystem. These are trivial conclusions made by Simakov after his theoretical studies of many years.

Diachronism of geological events in space is a main obstacle for elaboration of modern metering system based solely on their ages. Isochronous boundaries crosscutting paleosystems (the objection point by Simakov) are only the time templates, which do not chop the "system meat," as otherwise it would be very risky to visit a tailor. Of course, if necessary, we should use natural limits (boundaries of formations and zones) but this must not be the general principle of stratigraphy.

The defects of natural stratigraphy are not the simple consequences of its incompatibility with the correct metering procedure. They are interconnected with the specific character of stratigraphic investigations dealing with past events, which we can reconstruct with a significant allowance. This gives rise to subjective interpretations and endless unfruitful discussions. Using the biostratigraphic method, it is impossible to establish the ideal reference points determinable in geochronometry. However, the outlined "lighthouses" narrow considerably the sphere of subjective interpretations during determination of individual stratigraphic units and stratigraphic boundaries.

Unlike the natural stratigraphy, exochronometry does not consider characteristics of geological events, because it denies their adequate reflection in stratigraphic units, but some characteristic features of events should be considered here. The natural stratigraphy allows a wide range of subjective views on stratigraphic units and their boundaries of any rank, because it regards them as molds of certain geological (paleobiological) events. The so-called natural metrics are misleading, because we have no clear unambiguous understanding of evolutionary stages (events) and every event is actually recorded in many ways; to be exact, we deal always with a combination of interrelated events, boundaries between which do not coincide in time and deviate in space.

A wide spectrum of opinions is inevitable in this case. Each opinion appears to be correct, therefore, in accord with the principal definition that a stratigraphic unit and its boundaries must reflect a real event of appropriate rank. Discussions on significance of any event may be endless.

STRATIGRAPHIC CLASSIFICATION

Principles of stratigraphic procedure are inevitably reflected in stratigraphic classification (Table 2). Understanding of the conventional and the material in the unit content remains a central problem of modern stratigraphy, and this problem is dualistic in essence. There is a certain unanimity with respect to the conventional character of the Standard Scale units, but regional and local stratigraphic units used to be considered as records of particular stages in geological devel-

Table 2. Scheme of stratigraphic classification

Principal stratigraphic units					
Units of EXOCHRONOMETRY				Units of ENDOCHRONOMETRY	
Units of biochronometry			Units of radiochronometry, Ma	Units of lithostratigraphy	Units of litho- and tectonostratigraphy
Units	Units of Standard Scale	Regional units		Local	Regional
Phylochron (basic unit)	Zone	Local zone		Formation (basic unit)	Formation, group
Conventional unit	Stage, series, and others	Horizon		Litho- and tectonostratigraphic units	Complexes, tectonostratigraphic units

opment. This especially concerns the lithostratigraphic units and biostratigraphic zones, which signify in fact the heterogeneous composition of deposits and evolutionary diversification of fauna and flora.

Considering these units, we alienate the time characteristic of an object from its material content and accept the former as a time unit. Although any stratigraphic unit actually corresponds to a physical phenomenon, it is a conventional product of man's activity. The indicated correspondence is necessary to perform the metering, i.e., to reveal the geological time structure. Analysis of the metering procedure in stratigraphy shows that this correspondence is necessary only for the lowest basic units (biozones, formations).

Stratigraphic classification operates with two main sets of stratigraphic units: exochronometric and endochronometric. The exochronometric units, i.e., the main subdivisions of the Standard Scale and unified stratigraphic schemes, simultaneously represent biochronometric subdivisions and units of the absolute time scale. Biochronometric units are of different hierarchy: those of the Standard Scale are metering tools of a global stratigraphic time, and in regional and local stratigraphic schemes they depict successions of events.

As mentioned above, biochronometric units are diverse by nature. The basic (elementary) biostratigraphic units (zones, local zones) are based on of evolutionary peculiarities of concrete phyla (i.e., they are phylochrons), whereas of units of higher ranks (beginning from stages and horizon) are completely (or almost) unrelated to geological (paleontological) events of the corresponding significance.

Horizons reflect to some extent the geological development of regions, but they are biostratigraphic in essence, based on considerably autonomous evolution of fauna and flora. Like the stages, most of horizons represent a sum of elementary biostratigraphic units.

Therefore, we can distinguish biochronometric units of two ranks. Those of primary rank (biozones) represent elementary basic units (phylochrons) reflecting the evolution of elementary phyla (species, genera). As based on the lifetime of individual taxa, they exactly correspond to time ranges of elementary evolutionary events.

Biochronometric units of the second rank have scopes and boundaries chosen arbitrarily, though they are historically related to concrete strata. We can consider them as a sum of zones, i.e., as a set of elementary basic phylochrons. The time range of such a unit does not correspond to duration of any concrete geological or paleobiological event. The set of constituent zones (phylochrons), which determine boundaries, scope, and (to some extent) stratigraphic position of the higher-rank unit is itself a subject of agreement. The hierarchy of stratigraphic subdivisions, which consist of some or other set of basic units, is a matter of convenience in investigations (measurements).

Endochronometric units represent two hierarchic sets of similar content. Their specifics depends on principle used to distinguish the constituent stratigraphic units and on the kind of exochronometric scale, they are related to. The basic unit of the Phanerozoic endochronometry is a formation, boundaries of which are used in exochronometry to define chronostratigraphic (biostratigraphic) boundaries in accordance with the CFT principle. It should be emphasized that the main features characterizing a formation are the uniformity of its lithological composition and absence of internal unconformities and hiatuses. Lithostratigraphic units of a higher rank have no application in the Phanerozoic stratigraphy, because it is possible to use chronostratigraphic units in this case.

In the Precambrian stratigraphy employing the geochronological method, which is considerably less precise and yields fewer dates than the biostratigraphic method, lithostratigraphic units of higher rank (groups, complexes) are widely used in addition to formation. The leading role belongs to tectonostratigraphic levels, which are the main metering tools of the Precambrian scale. "Local Precambrian units of a high rank (groups) are recognized on the basis of the tectonic approach, i.e., by revealing unconformities that reflect tectonic movements and intrusive magmatic activity (magmatism) between the accumulation periods of supracrustal

deposits" (Stepanov and Mesezhnikov, 1979, p. 146). However, units of the formation rank are used in the Precambrian stratigraphy to trace stratigraphic boundaries. The main stratigraphic units are supplemented by a number of others, established by means of auxiliary particular methods (magnetostratigraphy, seismic stratigraphy, and others).

The scheme of stratigraphic classification suggested above is based on methods of distinguishing stratigraphic units. The corresponding features of stratigraphic units, which allow them to be the measuring tools, determine in what metering system the units function, or what a system they represent.

CONCLUSION

Development of stratigraphic methods, especially of zonal stratigraphy, prepared the ground enabling a correct metering procedure to function in stratigraphy. This is a consequence of "golden spike" concept and CFT principle introduced to define stratigraphic boundaries. Geochronometry is based on the external independent metering system, i.e., on exochronometry. Endochronometry, which is an intrinsic component of exochronometry, plays a subordinate role. Stratigraphic subdivisions can be regarded as conventional units of geological time. The recent step of stratigraphy toward geochronometry defines its role as a tool of measuring geological time. The whole structure of the geological time–space can be reconstructed only by geological mapping based on stratigraphy. Geological mapping and stratigraphy are inseparable components of a single method of graphic modeling of the geological development.

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The Boundary between Nemakit-Daldynian and Tommotian Stages (Vendian–Cambrian Systems) of Siberia

V. V. Khomentovsky and G. A. Karlova

Institute of Geology of Petroleum and Gas, Siberian Division, Russian Academy of Sciences, pr. Akademika Koptyuga 3, Novosibirsk, 630090 Russia

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Abstract—Distribution of small shelly fossils (SSF) in the Tommotian and upper Nemakit-Daldynian stages is revised in all main reference sections of the Siberian platform, particularly in its eastern facies zone. Correlation of coeval sections between each other and with the stratotype section of the transitional facies zone is substantiated in the northern and southeastern parts of the platform. As is shown, boundary between the indicated subdivisions is synchronous, marked by an extremely sharp change in abundance and diversity of SSF assemblages immediately above it. This fact and appearance of macrofaunal remains (archaeocyathids, brachiopods) imply that the Tommotian Stage exemplifies one of the most important phases in evolution of skeletal fauna. The boundary is also marked by a sharp negative excursion in the $\delta^{13}C$ curve. The obtained data on the lower boundary of the Tommotian Stage and its isotopic characteristic enable its global correlation and allow it to be accepted the lower boundary of the Cambrian System. The correlation method based on SSF assemblages and applied to different-facies Tommotian sequences is modified. It is important to analyze all local fossil assemblages and occurrence levels of younger taxa when substantiating boundaries of stratigraphic units instead of a limited number of guide species. Only this method enables recognition of age analogues of the Tommotian Stage subdivisions in the eastern and transitional zones.

Key words: stages, zones, small shelly fossils, zonal assemblages, facies zones, correlation, biostratigraphy, chemostratigraphy.

The problem of subdivision of the Nemakit-Daldynian stratigraphic unit and substantiation of its boundaries is of a principal significance for recognition and substantiation of lower boundaries of the Cambrian, Paleozoic, and Phanerozoic. In this article, we consider all available data on its upper boundary. The ambiguous interpretation of these data discredits important information on biostratigraphy of Siberia and substantiation of the Vendian–Cambrian units in the International stratigraphic scale. Savitskii (1962) was first to define oldest deposits of the Anabar shield, which bear small shelly fossils (SSF), as an autonomous Nemakit-Daldynian Horizon and referred it to the Cambrian. He established that the lower boundary of this unit coincides with a basal unconformity, and that the upper one corresponds to the base of variegated deposits (Medvezh'ya Formation) bearing archaeocyathids, brachiopods, and SSF of the *Oelandiella korobkovi–Anabarella plana* Zone. The latter zone was correlated with the Sunnagin and Kenyada horizons of the southern Siberian platform.

The different position of the Cambrian lower boundary in Siberia was proposed by I.T. Zhuravleva (Zelenov *et al.*, 1955) who subdivided the Zhurinian Substage of the Aldanian Stage into the Sunnagin, Kenyada, and Atdaban horizons based on archaeocyathid assemblages. Missarzhevskii and Rozanov (1963) correlated fossil assemblages from the basal part of the variegated sequence of the northern Siberian platform (the *Oelandiella korobkovi–Anabarella plana* Zone) with those from the Sunnagin–Kenyada deposits of the Aldan River basin and concluded that the mass appearance of skeletal fauna should be used to recognize the lower boundary of the Cambrian. It was noted that this boundary is also marked by the first occurrence of many SSF taxa and by changes in the phytolith assemblages. Taking this into consideration, Khomentovsky and Repina (1965) proposed to narrow the time-range of the Aldanian Stage and to correlate it with the upper Zhuranian Substage and the lower boundary of the Cambrian System with the base of the Variegated Formation (Sunnagin Horizon). Later on, Rozanov and Missarzhevskii (1966) excluded from the Aldanian Stage also its uppermost Atdabanian Horizon or Stage (Zuravleva, *et al.*, 1969) and termed the remainder as the Tommotian Stage.

The concept of the Tommotian Stage was eventually formulated in the monograph by Rozanov *et al.* (1969). They proposed a detailed biostratigraphic scheme that allowed this unit to be recognized beyond Siberia as principally distinct from the underlying sequence, the uppermost part of which was considered to be enclosing only rare problematics of the *Anabarites trisulcatus* Zone.

Works by Val'kov (1970, 1975) aggravated the situation with the Vendian–Cambrian biostratigraphy. He discovered the SSF assemblage of the *Oelandiella korobkovi–Anabarella plana* Zone in the upper part of the Manykai Formation, or Nemakit-Daldynian Horizon of the western Anabar region, and considered this zone as the pre-Tommotian in age. He correlated the newly distinguished *Allatheca cana* and *Allatheca anabarica* zones of the Medvezh'ya Formation with the *sunnaginicus* Zone. The Nemakit-Daldynian Horizon in its reduced range was renamed into the Manykai Horizon. Later on, Val'kov (1982, 1987) proposed to consider the interval between the *lata* and *cana* zones as an autonomous Khayalakhian Stage predating the Tommotian and representing, together with the underlying Manykaian Stage (*trisulcatus* Zone), the Chaburian, or Nemakit-Daldynian Superstage. In addition, he distinguished the *Spinulitheca billingsi* and *Spinulitheca rotunda* zones. Thus, the previously accepted zones of the complex substantiation were differentiated into zones based on different fossil groups—hyolithids, gastropods, and angustiocreids; moreover, these units were defined separately for the transitional and eastern facies zones of the Siberian platform.

Sequences from different zones and their boundaries were correlated based only on the so-called biostratigraphic markers—thin beds with zonal species. Val'kov had never explained how this is possible, when such beds are multiply repeated in the section, and how to use them for substantiation of boundaries. These aspects are discussed in detail below.

The diverse SSF assemblages reported by Val'kov from different sections of the eastern Siberian platform influenced significantly interpretations of many researchers. Casting doubts on the reliability of Val'kov's zonal scheme of Vendian–Cambrian fossil assemblages from northern and southern parts of the Siberian platform, Savitskii (1975) considered, nevertheless, the *plana* Zone as the pre-Tommotian in age. The ambiguous correlation of the *plana* Zone with the biostratigraphic zonal scale for the southern Siberian platform and doubts in reliable substantiation of the Nemakit-Daldynian unit boundaries (Savitskii, 1975; Savitskii *et al.*, 1980) allowed this unit to be classed with a regional horizon only.

Missarzhevskii who defined (together with Rozanov) the Tommotian Stage and substantiated its detailed SSF biozonation was also under the influence of Val'kov's ideas. He preferred the name Manykaian Stage instead of the Nemakit-Daldynian Horizon (or stage) and, like Val'kov, referred this unit to the Cambrian (Missarzhevskii, 1982,1983). Using the principle of Val'kov (1982), Missarzhevskii subdivided the Manykaian and Tommotian stages into seven (not five) interregional zones and defined, for the first time, the *Anabarella plana* Zone still identifying it with the *sunnaginicus* Zone. In distinction from his earlier works, Missarzhevskii, following Val'kov, placed the lower boundary of the Tommotian Stage first 15 (Missarzhevskii, 1982) and then 30 m (Missarzhevskii, 1989) above the base of the Medvezh'ya Formation of the Kotui River basin, and at least 9 m above the base of the Variegated Formation exposed near the Yudoma River mouth. After cessation of his activity in the Siberian platform, Missarzhevskii (1989), leaning mostly upon data by Val'kov (1987), also transferred the *plana* Zone (IV) from the Tommotian to Manykaian Stage and voluntarily united basal beds of the Tommotian Stage (zone V), which bear fossils of the *sunnaginicus* and *regularis* zones, into a single unit.

We have referred always and refer now the Nemakit-Daldynian unit to the Vendian System. Universally, this unit is overlain by the Tommotian Stage (*sunnaginicus* Zone). Its lower boundary corresponds to the base of a sequence with first oldest SSF assemblages (*trisulcatus* Zone). This paleontological substantiation allowed us to consider the unit as the Nemakit-Daldynian Stage (Khomentovsky, 1976).

Before the commencement of our work in the northern Siberian platform, we placed, following Val'kov, the lower boundary of the Tommotian Stage higher, inside the Medvezh'ya Formation (Khomentovsky, 1976). However, after works of 1979–1987 and biostratigraphic studies of the Vendian–Cambrian successions in the eastern facies zone of southeastern platform areas, we assured ourselves that interpretations by Val'kov and Missarzhevskii are incorrect. It became clear that variegated deposits, overlying the Vendian gray-colored carbonate rocks in the Siberian platform, are universally as old as at least the *sunnaginicus* Zone of the Tommotian Stage (Khomentovsky and Karlova, 1992, 1993). Rozanov (1992) also criticized schemes by Val'kov and Missarzhevskii who ignored, in his opinion, the influence of facies changes on composition of fossil assemblages.

When working in East Siberia, Semikhatov (Semikhatov *et al.*, 1970; Semikhatov and Serebryakov, 1983) correlated the lower boundary of the Cambrian with the base of the Variegated Formation and *sunnaginicus* Zone. Recently, he changed his view, however, and followed, together with his American colleagues (Kaufman *et al.*, 1996), the scheme by Val'kov and Missarzhevskii, placing the lower boundary of the Tommotian Stage in the Ust'-Kotuikan sequence of the western Anabar region as high as the uppermost part of the Medvezh'ya Formation. The formation was regarded as equivalent to the stratigraphic interval of a hiatus between the Variegated and Ust'-Yudoma formations in the Tommotian stratotype section (Dvortsy exposure at the Aldan River). Like Savitskii, these researchers correlate the lower boundary of the Nemakit-Daldynian Stage with the basal unconformity of the Manykai Formation and consider that stage as the basal Cambrian unit.

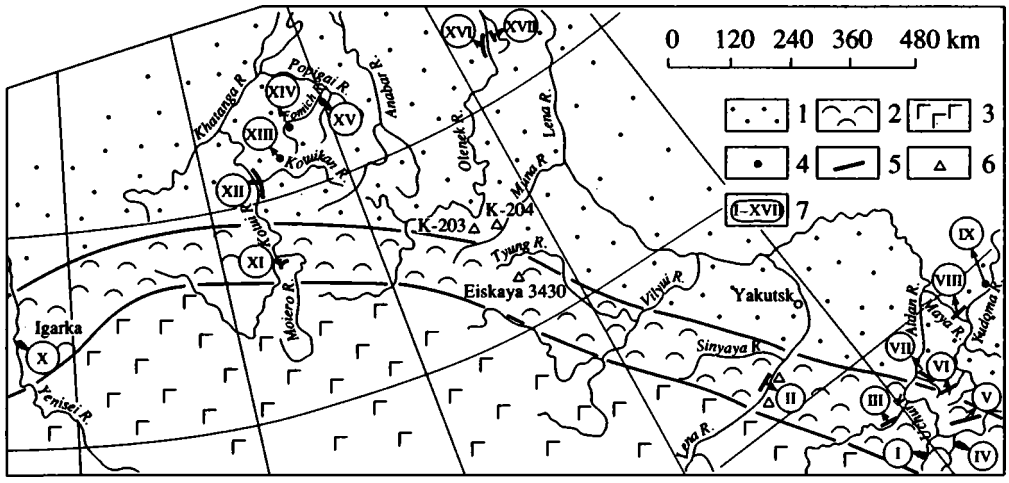


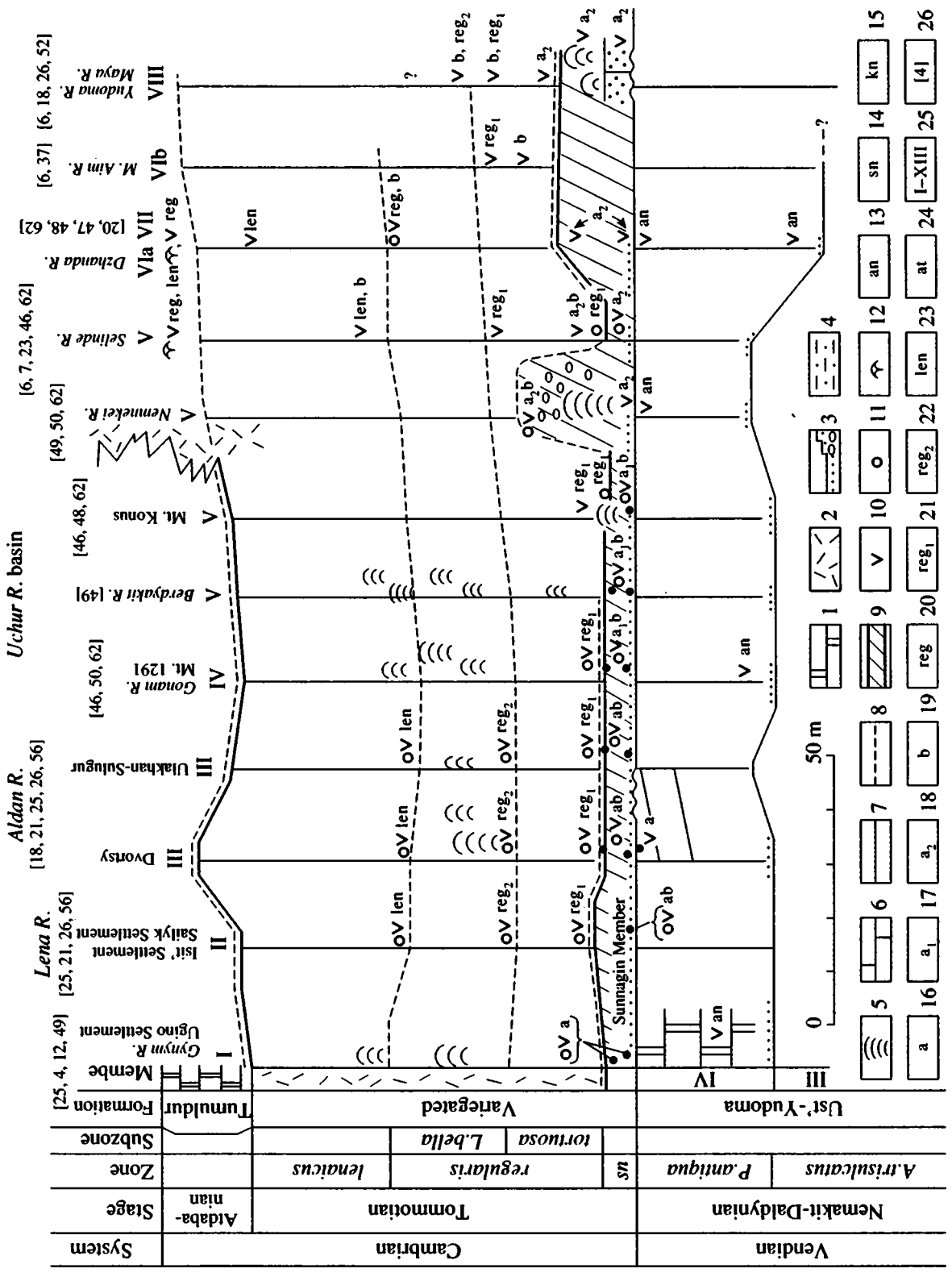
Fig. 1. Schematic zoning and location of sections discussed in the text: (1) eastern, (2) transitional, and (3) western facies zones; (4) exposure; (5) a series of exposures; (6) well; (7) exposure numbers also indicated in Figs. 2 and 3.

LITHOSTRATIGRAPHIC CHARACTERIZATION OF THE TOMMOTIAN LOWER BOUNDARY IN SOUTHEASTERN SIBERIAN PLATFORM

As is shown above, the controversial positioning and substantiation of the Cambrian lower boundary in Siberia is mainly a consequence of problematic correlation between Vendian–Cambrian sections in different facies zones of the Siberian platform. It is well known that its southwestern part was occupied at that time by a spacious lagoon that accumulated deposits of the western or Olekma facies zone (Fig. 1). The eastern and northern parts of the platform were covered by a spacious shelf sea basin now regarded as the eastern or Yudoma–Olenek facies zone. Both basins were separated by a relatively narrow elevation corresponding to the transitional or Sinian-Anabar facies zone (Savitskii, 1972; Khomentovsky and Repina, 1965; *Yarusnoe raschlenenie...*, 1984). The elevated water salinity in the lagoon was unfavorable for biota evolution, and paleontological remains are scarce in the corresponding zone. To the contrary, the shallow-water rise flooded by sun-heated fresh sea water flowing in the lagoon favored development of animal and plant organisms. As a result, precisely this transitional facies zone hosts the biostratigraphic standard of the Lower Cambrian (Rozanov *et al.*, 1969). Biocoenoses of the open eastern sea of the Tommotian time are characterized by a wide distribution rather than by abundance of fossil taxa (Khomentovsky and Karlova, 1992, 1993; Rozanov, 1992). The biostratigraphic standard of the transitional zone is variably correlated with sections of the eastern zone. The widely accepted standpoint is that the diverse assemblages are diachronic in the eastern and transitional zones. It is assumed that habitats favorable for Metazoa first appeared in the former, and that the fossil assemblage of the *sunnaginicus* Zone in the stratotype section of the Tommotian Stage appeared as a result of westward migration of younger Metazoa communities (Savitskii, 1972; Val'kov, 1987; Missarzhevskii, 1989). Other researchers (Kaufman *et al.*, 1996; Landing, 1994) who assume a large hiatus at the base of the Variegated Formation arrived at the same conclusion.

Nevertheless, real relationships between the Variegated and Ust'-Yudoma formations disagree with such a model. In terms of sedimentogenesis, the Ust'-Yudoma Formation is quite close to the Variegated one. It includes facies elements, which point to the commencement of lagoon formation at the salt accumulation time. In the eastern direction, the Ust'-Yudoma dolomites grade into limestones (Semikhatov *et al.*, 1970). The barrier that separated the basins already existed, as it is evident from a significant reduction of the Ust'-Yudoma Formation thickness from 180–200 to 120 m in the area between the Nemnekei and Selinde rivers, the right tributaries of the Uchur River (Khomentovsky and Karlova, 1993, 1994a).

Extremely important for recognition of the lower boundary of the Variegated Formation and Tommotian Stage in the southeastern Siberian platform is a lithological marker that is termed the Sunnagin Member or Horizon (Fig. 2) (Zelenov *et al.*, 1955). The marker is a basal unit of the Variegated Formation that differs from its other beds in gray–green color, abundance of glauconite, and in a significant admixture of detrital material. The Sunnagin Member is traceable over the entire distribution area of the Variegated Formation from its westernmost outcrops near the Tommot Settlement at the Aldan River to easternmost sections in the Yudoma–Maya depression. Almost everywhere, the member bears erosion signs at its base. Nevertheless, its isochronism is evident from the persistent thickness of Member IV of the Ust'-Yudoma Formation (23–28 m), on the one hand, and Sunnagin member proper (4–8 m), on the other. The factual data imply that the hiatus mentioned above is insignificant, having no tendency to increase westward. Within the westerly Berezovo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Nemakit-Daldynian
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Aldaba-nian
Stage
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L. bella
tortuosa
Subzone
Zone
Ust'-Yudoma
Varegated
Tumuldur
Formation
[25, 4, 12, 49]
Lena R.
[25, 21, 26, 56]
Aldan R.
[18, 21, 25, 26, 56]
Uchur R. basin
[6, 7, 23, 46, 62]
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Member
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Gynym R.
Isit' Settlement
Salyk Settlement
Dvorsy
Ulakhan-Sulugur
Gonam R.
Mt. 1291
Berdyakit R. [49]
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depression, where the Variegated Formation is almost completely replaced by gray-colored dolomites, the gradual transition at the base complicates determination of the lower boundary. In western sections of the Aldan River area, the hiatus is locally missing at all. The gradual transition between the Ust'-Yudoma and Variegated formations is established in the middle part of the stratotype Tommotian section of the Dvortsy exposure (Fedorov *et al.*, 1992; Khomentovsky and Karlova, 1992) and at the Nemnekei and Dzhanda rivers (Khomentovsky and Karlova, 1989, 1991). In the area between the Dvortsy and Ulakhan-Sulugur exposures, several intermediate sections with the marker member show that the Ust'-Yudoma Formation thickness gradually decreases in the eastern direction, because its upper five meters are wedging out (Semikhatov *et al.*, 1970; Avdeeva *et al.*, 1983; Khomentovsky *et al.*, 1990; Khomentovsky and Karlova, 1992, 1993). The most significant hiatus at the base of the Variegated Formation is registered near the Yudoma River mouth and in the Kyllakh Ridge located to the east and north of the above-mentioned exposures, respectively (Mezentsev *et al.*, 1978; Starnikov *et al.*, 1983; Sukhorukov, 1984). Some researchers exaggerated significance of this hiatus directly correlating sections, in which the Variegated Formation overlies the dark gray limestone member of the Ust'-Yudoma Formation, with those, where the it rests on the upper light dolomites of the Ust'-Yudoma Formation (Starnikov *et al.*, 1983; Sukhorukov, 1984). However, sections of the first type are located far to the north, and total thickness of the Ust'-Yudoma Formation sharply decreases in this direction. Simultaneously, composition of SSF sampled in a bed-by-bed manner (Starnikov *et al.*, 1983; Sukhorukov, 1984) indicate that beds with SSF of the *P. antiqua* Zone, which are universally present in sections of the transitional facies zone, are truncated in the north (Khomentovsky and Karlova, 1992, 1993).

A significant hiatus at the base of the Variegated Formation is locally registered in the Kyllakh Ridge only. Near the Ulakh-Ytyga Creek mouth, upstream of the Yudoma River, sections are almost continuous (Khomentovsky *et al.*, 1972; Semikhatov and Serebryakov, 1983). The anomalous magnitude of hiatus is probably related to displacement along the Nel'kan fault, which divided the Uchur–Maya plate and Yudoma–Maya depression at that time. It is also probable that hiatuses originated owing to intense hydrodynamics at local elevations, whereas sedimentation in deeper and calmer setting was continuous (Khomentovsky and Karlova, 1992). Thus, geological data evidence against diachronism of the lower part of the Variegated Formation in the transitional and eastern facies zones.

PALEONTOLOGICAL SUBSTANTIATION OF THE NEMAKIT-DALDYNIAN–TOMMOTIAN BOUNDARY IN THE SOUTH OF THE PLATFORM

Two aspects should be taken into consideration when discussing position of the Nemakit-Daldynian–Tommotian boundary in the southern region of Siberian platform. (1) Orthostratigraphic fossil groups of the Atdabanian and Tommotian stages are trilobite + archaeocyathid and archaeocyathid + brachiopod assemblages, respectively, but the Nemakit-Daldynian Stage is established and subdivided based on integral SSF assemblages. (2) It is more difficult to use SSF assemblages of *antiqua* and *sunnaginicus* zones in stratigraphic practice than orthostratigraphic macrofaunal groups (Khomentovsky and Karlova, 1992, 1994a). Correlation of Vendian–Cambrian sections in different facies zones is a main objective of this work.

Those who believe that classical sections of the Tommotian Stage in middle courses of the Aldan River have a large-scale basal hiatus note first of all the remarkable difference in composition of SSF assemblages from the Sunnagin Member of the Variegated Formation and from the uppermost Member IV of the Ust'-Yudoma Formation. The SSF assemblage from the former includes 54 taxa whereas the latter yields only six forms: *Chancelloria, Cambrotubulus decurvatus, Anabarites trisulcatus, Halkieria sacciformis, Hyolithellus vladimirovae*, and *H.* sp. A half of the latter is inherited from the older *trisulcatus* Zone (Khomentovsky and Karlova, 1993, 1994). Indeed, the difference (47 taxa) is significant. Nevertheless, this argument is invalid because the elevation that separated the eastern and western facies zones at the time of the *antiqua* Zone was located, as is shown above, far to the east of the Dvortsy–Ulakhan-Sulugur area that corresponded to elevation of the initial Tommotian time.

Fig. 2. Correlated sections and fossil assemblages of the southeastern Siberian platform: (1) gray dolomites; (2) varicolored clayey–carbonate rocks; (3) basal layers; (4) terrigenous rocks; (5) bioherms; (6) limestones; (7) lithostratigraphic boundaries; (8) biostratigraphic boundaries; (9) marker members; (10) SSF; (11) archaeocyathids; (12) trilobites; (13) *antiqua* and (14) *sunnaginicus* zones; (15) Kyndyn member; (16) SSF of the *sunnaginicus* Zone in the transitional zone; (17) ditto plus insignificant amount of SSF from the eastern facies zone; (18) ditto plus abundant SSF appearing in the *sunnaginicus* Zone of the eastern facies zone; (19) first occurrence of "determining" taxa of the *sunnaginicus* Zone; (20) *D. regularis* Zone; (21) *L. tortuosa* Subzone; (22) *L. bella* Subzone; (23) *D. lenaicus* Zone; (24) Atdabanian Stage; (25) numbers of exposures I–XIII indicated in Fig. 1; (26) number of selections in Fig. 1. References: [4] Val'kov, 1978, [6] Val'kov, 1987, [7] Voronova *et al.*, 1983, [12] Korshunov, 1972, [18] Missarzhevskii, 1989, [20] Pel'man *et al.*, 1990, [21] *Putevoditel'...*, 1973, [23] Repina *et al.*, 1988, [25] Rozanov *et al.*, 1969, [26] Rozanov *et al.*, 1992, [[37] Sysoev, 1972, [46] Khomentovsky and Karlova, 1986, [47] Khomentovsky and Karlova, 1989, [48] Khomentovsky *et al.*, 1990, [49] Khomentovsky and Karlova, 1991, [50] Khomentovsky and Karlova, 1992, [52] Khomentovsky and Karlova, 1994a, [56] *Yarusnoe raschlenenie...*, 1984, [62] Khomentovsky and Karlova, 1993.

Therefore, the impoverished assemblage of the *antiqua* Zone characterizes, in fact, the western zone. However, if we take fossils characterizing the elevation area of the corresponding time (Nemnekei and Selinde rivers) and neighboring sections (Gonam River, Mt. Konus), the number of taxa increases in the SSF assemblage to 24 (Khomentovsky and Karlova, 1993). The difference in 23 taxa, as compared to the *sunnaginicus* Zone assemblage, is important of course from the standpoint of evolution, but it does not necessarily mean a substantial hiatus in sedimentation. Slightly eastward (Dzhanda River), the SSF assemblage of the *antiqua* Zone is more diverse, as it includes 10 additional species characteristic of the eastern zone, thus being comparable in diversity with that of the *sunnaginicus* Zone in classical sections of the Aldan River. Most important species of this assemblage are *Purella antiqua, P. cristata, Latouchella korobkovi, Tiksitheca licis, Archaeospira regularis, Ladatheca annae, Loculitheca anulata, Halkieria sacciformis, H. longa, H. amorphe, Igorella* sp., *Barskovia* sp., *Helcionella* sp., *Siphogonuchites triangulaspis, Hyolithellus vladimirovae*, and *Markuelia prima.*

We cited complete lists of SSF taxa from the *sunnaginicus* Zone (Khomentovsky and Karlova, 1993, 1994). Here, we mention only about 20 taxa, which are most important for correlation of different-facies sections. These are *Allatheca concinna, Exilitheca multa, Spinulitheca billingsi, Conotheca mammilata, Turcutheca crasseocochlia, Edgetheca aldanica, Aldanella rozanovi, A. attleborensis, A. crassa, A. utchurica, Barskovia hemisymmetrica, Bemella jacutica, Isitiella inprocera, Igorella ungulata, Watsonella sibirica, Torellella curvae, T. lentiformis, Hyolithellus tenuis, Tommotia kozlowskii, T. admiranda*, and *Camenella garbowskae.*

Another argument in favor of significant hiatus at the base of the Variegated Formation in the transitional facies zone consists in the statement that the complete set of SSF taxa of the *sunnaginicus* Zone appears here immediately above its base, whereas in the eastern facies zone some of them appear significantly higher in the section. Researchers who admit the older age of Variegated Formation in the eastern facies zone propose to distinguish here analogues of subdivisions defined in the Tommotian stratotype section, using occurrence of such taxa. Adherents of this standpoint (Val'kov, 1987; Missarzhevskii, 1989; Kaufman *et al.*, 1996) published different lists of "determining" taxa, and all of them listed the following forms: *Edgetheca aldanica, Allatheca concinna, Torellella curvae, Watsonella sibirica, Tommotia admiranda*, and *T. kozlowskii.* The lists often include also taxa that appear near the base of the *regularis* Zone and are characteristic of the latter in the classical sections of the Aldan River middle courses: *Lapworthella tortuosa, Allatheca corrugata, Tommotia plana.* Sometimes (Val'kov, 1987), even *Burites distortus* and others are considered to occur at the base of the Tommotian Stage. Therefore, we exclude the mentioned taxa from the list characterizing the *sunnaginicus* Zone. To test whether the position of lower Tommotian boundary can be correctly established based on the "guide" taxa only, let us consider how the Tommotian SSF assemblages change their composition in facies zones located close but away from the transitional elevation. West of the line Ulakhan-Sulugur–Dvortsy, the transitional facies zone stretched up to the Tolba River mouth (*Yarusnoe raschlenenie...*, 1984), however, already 30 km upstream the Aldan River, near the Ugino Settlement, the fossil assemblage from the Sunnagin Member becomes substantially impoverished. In addition to several forms inherited from the Ust'-Yudoma Formation, this member yields here only 8 taxa typical of the *sunnaginicus* Zone: *Edgetheca aldanica, Coleolella billingsi, Turcutheca crasseocochlia, Coleolus trigonus, Aldanella rozanovi, Hyolithellus tenuis, Torellella lentiformis, T. curvae*, and *Tumulduria incomperta* (Rozanov *et al.*, 1969). In the Gynym River middle courses, slightly further westward of the elevation (Figs. 1, 2), the typical Sunnagin Member yields, in addition to several forms passing from the Ust'-Yudoma Formation, only *Torellella lentiformis, Sunnaginia* sp., and *Spinulitheca* sp. This assemblage is lacking any "determining" taxon, but the coexisting brachiopods (*Aldanotreta* sp.) and archaeocyathids of the *sunnaginicus* Zone (Korshunov, 1972; Val'kov, 1978; Khomentovsky and Karlova, 1991) unambiguously imply its Tommotian age. Thus, the absence of "determining" taxa cannot mean the age difference between beds with SSF assemblages slightly different from that of the *sunnaginicus* Zone stratotype. In addition, when the impoverished SSF assemblage of the *sunnaginicus* Zone from sections located near the western facies zone is compared with that of the *antiqua* Zone from the Dvortsy section, which occupied a similar facies position at that time, the difference that was used to substantiate a large-scale hiatus between these zones becomes negligible.

Changes in paleontological characteristics of the *sunnaginicus* Zone are also characteristic of the transitional facies zone east of the elevation. The exposure at the summit of a mount (abs. mark 1291 m) at the right side of the Gonam River is a geographically closest to the past elevation (Figs. 1, 2). Here, the Vendian–Cambrian section is almost identical to that exposed in the Dvortsy locality (Khomentovsky *et al.*, 1983; Khomentovsky and Karlova, 1986, 1992, 1993). The Variegated Formation begins here with the typical Sunnagin Member about 5 m thick. Its uppermost part encloses archaeocyathid bioherms rising up by 3–7 m into red beds. The lower part of bioherms is composed of archaeocyathids of the *sunnaginicus* Zone, whereas the upper one yields fossils of the *D. regularis* Zone (Rozanov *et al.*, 1969; Khomentovsky *et al.*, 1983; Khomentovsky and Karlova, 1992). The *sunnaginicus* Zone yields 33 SSF taxa, including seven inherited from the *antiqua* Zone. The assemblage contains main reference forms characteristic of the *sunnaginicus* Zone in the Dvortsy section, including the "determining" taxa

(Khomentovsky and Karlova, 1993). Of a particular importance is the appearance of taxa that are missing from classical Tommotian sections, but common in the eastern facies zone: *Loculitheca anulata, Siphogonuchites* sp., *Halkieria operculus*, and *Sachites proboscideus.*

Eastward, the Sunnagin Member, isochronism of which is substantiated by all geological data and by archaeocyathids of the *sunnaginicus* Zone, is traceable in the almost continuous well-exposed sections of the Ketkap Ridge, the Aldan–Uchur water divide, extending to upper courses of the Nemnekei River (Figs. 1, 2) (Khomentovsky and Karlova, 1986, 1991, 1992). The trend of changes in the SSF composition that was noted at the mount with altitude of 1291 m is also characteristic of this area. In upper courses of the Berdyakit River, the SSF assemblage consists of 11 taxa, including tommotids *Camenella garbowskae* peculiar of the *sunnaginicus* Zone in the stratotype section, while the SSF assemblage from the eastern facies zone includes also *Aldanella crassa* and *Ovalitheca* sp. (Khomentovsky and Karlova, 1991). The undoubted Sunnagin Member exposed at the Mt. Konus slope yields 14 characteristic species that are missing from the *antiqua* Zone. Tommotids are represented only by *Tommotia* sp. Taxa characteristic of the eastern facies zone and unknown in the western one are *Yochelcionella stilifera, Halkieria processa*, and *Hyolithellus tschuscunensis*, the latter widespread in the *sunnaginicus* Zone (Khomentovsky and Karlova, 1986, 1991, 1993). In the Nemnekei River section located 20 km east of Mt. Konus, the uppermost part of the light dolomite member of the Ust'-Yudoma Formation is marked by microkarst features and by occurrence of flat dolomite pebbles. Previously, we suspected that these are indications of hiatus at the base of the Variegated Formation. It appeared, however, that rare SSF are represented here only by forms characteristic of the *antiqua* Zone (Khomentovsky and Karlova, 1991, 1992, 1993). Overlying yellowish gray Sunnagin member about 4–5 m thick is mainly composed of dolomitic sandstones. They contain, along with forms of the *sunnaginicus* Zone typical of its stratotype, species characteristic of the eastern sections: *Ovalitheca mongolica, Kotuitheca curta, Latuochella maidipingensis, Tannuella gracilis*, and *Pollicina* sp. Similarly to other sections of the transitional facies zone, the upper part of the Sunnagin Member is represented by biohermal buildups. The latter are 15–20 m high together with products of their destruction and composed of calcareous algae, instead of archaeocyathids (Khomentovsky and Karlova, 1991, 1992, 1993). The lower 3 m of the overlying Variegated Formation enclose first *Allatheca concinna, Aldanella attleborensis, A. rozanovi, Conotheca mammilata*, and *Watsonella sibirica*, which are typical of the *sunnaginicus* Zone in the transitional sections, and diverse species abundant in the eastern zone: *Nikatheca digna, Igorella hamata, Kotuitheca* sp., *Emarginoconus mirus, Triplicanthella disdoma, Crossbitheca* sp., and *Mellopegma* sp. The Tommotian age of this level is confirmed by fragments of archaeocyathids observed in preparations. Thus, the *sunnaginicus* Zone of the Nemnekei River section includes, in equal proportions, taxa of both the transitional and eastern facies zones.

In the western areas of the eastern facies zone, the Sunnagin member is traceable sufficiently well. In the upper part of the Selinde River (50 km east of Mt. Konus), the Ust'-Yudoma dolomites are overlain with a hiatus, like in the Aldan River section, by yellowish gray calcareous dolomites and dolomitic sandstones 4–6 m thick. Depressions in the Ust'-Yudoma Formation surface are filled with coarse-grained sandstones enriched in glauconite (Khomentovsky and Karlova, 1986). Simultaneously, the paleontological characteristic of this member is ambiguous. Most researchers (Korshunov *et al.*, 1969; Voronova *et al.*, 1983; Khomentovsky and Karlova, 1986, 1992, 1993; Repina *et al.*, 1988) believe that it should be referred entirely to the *sunnaginicus* Zone. Val'kov (1987) argued, however, that the Sunnagin Member as well as 23 m of the overlying Variegated Formation are pre-Tommotian in age, and Missarzhevskii (1989) shared his opinion. Such a conflicting interpretation can be explained by difference between the SSF assemblages from the Selinde River section and those from the Dvortsy and Ulakhan-Sulugur sections. Interpretations by Val'kov and Missarzhevskii are discredited by our finds of archaeocyathid fragments 2.7 m above the base of the Variegated Formation and by data that rocks located 5 m above the Sunnagin Member yield *Nochoroicyathus mirabilis* and *Cambrocyathellus tschuranicus* (Korshunov *et al.*, 1969) characteristic of the *D. regularis* Zone. Accordingly, it can be concluded that, similarly to the western sections, the Sunnagin Member in the area under discussion is not older than the *sunnaginicus* Zone. The data on SSF distribution contradict the aforementioned standpoint as well. The assemblage from the Sunnangin Member is known to include 71 taxa (Repina *et al.*, 1988; Khomentovsky and Karlova, 1993), 16 of which are inherited from the *antiqua* Zone, and 18 forms are typical of classical Sunnagin sections at the Aldan River. Simultaneously, the SSF assemblage from the Selinde River section becomes to be dominated by taxa characteristic of the basal part of the Tommotian Stage in the eastern facies zone. Nine of them have been mentioned among taxa of the intermediate sections, whereas 24 species are newcomers in the section under consideration. These are *Turcutheca rugata, T. cotuiensis, Spinulitheca rotunda, Latouchella adelocosma, Nomgoliella sinistrivolubilis, Tannuella gracilis, Stenothecoides, Gastreochrea viva, Rostraconus sinensis, Yangzespira regularis, Anabarithellus hexasulcatus, Pseudoortotheca* sp., *P. costata, Obtusoconus multicostatus, Prooneotodes?, Kugdatheca voluta, Xianfengella* sp., *Paraaldanella kotujica, Kraikhania* sp., *Mariochrea sinuosa, Fomitchella infundibuliformis, F. acinaciformis, Watsonella varensalensis*, and *Selindeochrea tecta.*

Thus, judging from the SSF composition, the Sunnagin Member in the Selinde River section belongs undoubtedly to the eastern facies zone. Species *Aldanella rozanovi, A. utchurica, A. attleborensis, Hyolithellus tenuis, Tommotia* sp., and others occurring its base upward and typical of the *sunnaginicus* Zone (Repina *et al.*, 1988; Khomentovsky and Karlova, 1993), as well as the data on archaeocyathids, indicate the Tommotian age of the member. The standpoint of Val'kov (1987) that overlying 10 m of the Variegated Formation should be referred in the Selinde River section to his pre-Tommotian *Spinulitheca rotunda–Anabarella plana* Zone is invalid, because both index species of this zone coexist with *Allatheca cana* and typical Sunnagin taxa *Spinulitheca billingsi, Exilitheca multa, Allatheca concinna*, and *Conotheca mammilata* in this interval (Repina *et al.*, 1988; Khomentovsky and Karlova, 1993). Val'kov believed that the Tommotian Stage in the Selinde section begins 28 m above the base of the Variegated Formation. However, this level yields abundant SSF species of the *regularis* Zone: *Allatheca corrugata, Notabilites simplex*, and *Oblisicornus dupleconcavus*. Their appearance is lagged behind coeval archaeocyathids for a period corresponding to formation of the 20-m-thick Variegated Formation. In a similar manner, *Torellella lentiformis* and *Camenella garbowskae* typical of the *sunnaginicus* Zone are recorded in this section almost simultaneously with SSF forms of the *lenaicus* Zone, and several taxa of the *regularis* and *lenaicus* zones appear only at the level with Atdabanian trilobites (Repina *et al.*, 1988).

The Vendian–Cambrian section of the Dzhanda River (right tributary of the Aldan River) contains the most representative SSF assemblage of the eastern facies zone (Figs. 1, 2). Three complete sections of this interval have been studied here in: (1) Exposure VIa, 2 km downstream of the Olom Creek mouth (75 km above the Dzhanda River mouth; (2) Exposure VIIa (25 km upstream of the mouth); and (3) Exposure VIIb, 15 km upstream of the mouth. Like everywhere, the Sunnagin Member is composed of greenish and yellowish gray dolomites crowned by a marker bed of white biohermal limestones (Fig. 2). Its thickness is 13 m (Khomentovsky and Karlova, 1989; Khomentovsky *et al.*, 1990). The member conformably rests upon the Ust'-Yudoma dolomites, and the contact is marked by interbeds of dolomitic, glauconite-bearing sandstones up to 10 cm thick, though rare lentils enriched in glauconite grains occur lower as well. The overlying red beds of the Variegated Formation are exposed fragmentarily in the Dzhanda River sections. There are excellent outcrops of Vendian–Cambrian rocks along the Ende River, between exposures VIa and VIIa, but the Sunnagin Member is unexposed here. This leads to erroneous correlation of sections, for instance, to overestimated thickness of bed corresponding to the *sunnaginicus* Zone. We (Khomentovsky and Karlova, 1989) also established that in Exposure VIIa (D_{8715} in Pel'man *et al.*, 1990), the interval from the Ust'-Yudoma Formation roof to beds referred by Pel'man with colleagues to the *regularis* Zone equals to 15, but not to 40 m.

The find of *Aldanella rozanovi* and *A. attleborensis* 30 m below the base of the Variegated Formation in the southern continuation of Exposure VIIb, which was mentioned in the work by Pel'man *et al.* (1990), also casts doubts. Representatives of genus *Aldanella* were never found in any outcrop of the Ust'-Yudoma Formation at the Dzhanda River. Only *Anabarella plana* and *Hyolithellus tschuskunensis* are recorded in these sections along with *Anabarites* forms. We traced the bed of clayey dolomites with these fossils over the entire Exposure VIIb, upstream and downstream of the Yuesee-Kymystan Creek, the right tributary of the Dzhanda River. It was established that this bed corresponds to the basal part of Member IV of the Ust'-Yudoma Formation in our scheme, and that beds in this area distinctly dip northward and are faulted (Khomentovsky and Karlova, 1989). During next visit to the Dzhanda River section (Khomentovsky *et al.*, 1990), we already knew data reported by Pel'man and his colleagues and paid a special attention to sampling of basal part of Member IV. Several SSF forms of the *antiqua* Zone sampled additionally in some localities along the strike are *Tiksitheca licis, Kotuithes biplicatus, Purella* sp., *Paleosulcachites* sp., *Protohertzina anabarica*, and *P. unguliformis* (Khomentovsky *et al.*, 1990). Nevertheless, we failed to discover either *Aldanella* or other Tommotian forms. We do not exclude that Pel'man with colleagues ignored a complicated structure of Exposure VIIb and found *Aldanella* remains in rocks visually similar to those of the Sunnagin Member. Near the roof of the Ust'-Yudoma Formation, the SSF assemblage of the *antiqua* zone includes the following additional taxa: *Loculitheca beata, Selindeochrea* sp., *Latouchella maidipingensis, L. korobkovi, Purella antiqua, P. cristata, Archaeospira regularis*, and *Lophotheca socialis* (Khomentovsky *et al.*, 1990; Khomentovsky and Karlova, 1993), but *Aldanella* forms are absent at this level as well. As usually, the diversity of the SSF assemblage from the Sunnagin Member sharply increases to 55 taxa at this site, apart from new and endemic forms (Khomentovsky *et al.*, 1990; Pel'man *et al.*, 1990; Khomentovsky and Karlova, 1993). Among them, 15 species are from the *antiqua* Zone, and 17 taxa are typical of the *sunnaginicus* Zone in the Dvortsy and Ulakhan-Sulugur sections, including brachiopods *Aldanotreta* sp. and SSF species *Igorella ungulata, Aldanella utchurica, Hyolithellus tenuis, Barskovia hemisymmetrica, Watsonella sibirica, Edgetheca* sp., *Bemella jacutica, B. septata*, and *B. parula* (Pel'man *et al.*, 1990; Khomentovsky and Karlova, 1993). Nineteen taxa, many of which were mentioned among species of the intermediate sections, are typical of the eastern facies zone. Other eastern taxa are represented by *Kotuithes biplicatus, Fomitchella acuticostata, Nikatheca* sp., *Rostraconus sinensis, Kotuikanites vallatus*, and *Granoconus strem-*

atus. Only these beds yielded *Aldanella rozanovi, A. attleborensis*, and *A. crassa*.

The Dzhanda River sections demonstrate a tendency similar to that established in adjacent sections: the appearance of taxa characteristic of the transitional facies zone is lagged behind that of the eastern zone SSF. For instance, *Allatheca concinna, Turcuheca* sp., *Jacutiolituus fusiformis, Torellella lentiformis, T. curvae*, first *Tommotia* spp., *Coleoloides tipicalis*, and *C. trigeminatus* appear only at the level of archaeocyathids of the *D. regularis* Zone, whereas in the classical Tommotian sections at the Aldan River, they occur already at the base of the Sunnagin Member. *Burites distortus* and *B. erum* characteristic of the latter zone first occur in the Dzhanda River sections only above the first occurrence level of trilobites of the Atdabanian Stage (Pel'man *et al.*, 1990).

The exact position of the boundary between *sunnaginicus* and *regularis* zones in the Dzhanda River section is not established, because the 30-m-thick boundary interval between them lacks guide forms of the last zone (Pel'man *et al.*, 1990). However, in upper courses of the Malyi Aim River (Exposure VIb, Figs. 1, 2), 30 km east of the Dzhanda River locality, this interval of uncertainty is narrower (Sysoev, 1972). The lower 8 m of red beds of the Variegated Formation overlying the Sunnagin Member contain here SSF species typical of the *sunnaginicus* Zone: *Spinulitheca billingsi, Allatheca concinna*, and *Exilitheca multa*. Four meters higher, rocks yield the SSF assemblage of the *regularis* Zone: *Allatheca corrugata, Notabilitus simplex*, and *Oblisicornus tetraconcavus*. These finds confirm the geologically substantiated conclusion that the boundary between these zones corresponds to the roof of the Sunnagin Member, like in all other considered sections.

To substantiate the pre-Tommotian age of the basal part of the Variegated Formation in this area, Val'kov (1978, 1982, 1987) used sections at the Malyi Aim River located directly below its mouth. However, according to our data, the Ust'-Yudoma Formation is separated there from red beds of the Variegated Formation by a fault that truncates completely the Sunnagin Member. The oldest layers of the Variegated Formation exposed in this section yield the following taxa of the *sunnaginicus* Zone: *Spinulitheca billingsi, Fomitchella infundibuliformis, F. acinaciformis, Anabarithellus hexasulcatus, Aldanella attleborensis*, and *A. utchurica*. Therefore, despite presence of some forms of the *antiqua* Zone, these rocks cannot be considered as the pre-Tommotian in age. The overlying bed contains, along with guide forms of the *sunnaginicus* Zone of transitional sections (*Exilitheca multa, Sachites proboscideus, Camenella garbowskae, Aldanella rozanovi, Sunnaginia* sp.), an additional eastern taxon *Allatheca cana*, which has been considered, for the unknown reason, as a guide taxon of new, synonymous pre-Tommotian zone (Val'kov, 1982, 1987). It is noteworthy that directly higher, species *Allatheca corrugata, Gracilitheca bayonet, Eonovitatus superbus*, and *Tchuranitheca simplicis* characteristic of the *regularis* Zone have been detected (Val'kov, 1978).

Further to the east, in lower courses of the Yudoma River and adjacent area of the Maya River basin (Figs. 1, 2), the hiatus at the base of the Variegated Formation is, as shown, of the maximum duration, and the basal part of the Sunnagin Member includes thicker beds (up to 3–4 m) of dolomitic sandstones and siltstones (products of reworking of Ust-Yudoma dolomites). Like in the sections at the Nemnekei and Selinde rivers, the upper half of the member is composed of light biohermal limestones alternating with platy greenish gray marls and clayey limestones, all commonly bearing glauconite. The thickness of the Sunnagin Member is 10–12 m. The direct contact between the Ust'-Yudoma dolomites and overlying Variegated Formation is exposed about 25 km upstream of the Yudoma River mouth, where it is marked by a slight angular unconformity (Khomentovsky and Karlova, 1994b). Red beds of the Variegated Formation rest on the Sunnagin Member conformably, with a gradual transition. In this and other exposure located at the Maya River 27 km downstream of the Yudoma River mouth, the succession of boundary beds between the Ust' Yudoma and Variegated formations is beyond question, and the list of fossils from our collection (Khomentovsky and Karlova, 1994b) can be supplemented with those sampled by Val'kov (1982, 1987) and Missarzhevskii (1989). The SSF assemblage from the lower part of the Sunnagin Member is extremely impoverished. In addition to several forms, which transit from the *antiqua* Zone, it includes typical species of the member, such as *Aldanella rozanovi, A. attleborensis, A. utchurica*, and *Spinulitheca rotunda*, the taxon occurring at the same level in western sections. In biohermal and plate greenish gray limestones, the assemblage becomes more diverse. It includes many transit species characteristic of the Ust'-Yudoma Formation and also known from the Variegated Formation (*Anabarella plana, Ladatheca dorsocava, L. annae, Oelandiella korobkovi, Tiksitheca licis*, and others). These coexist with typical forms of the *sunnaginicus* Zone: all *Aldanella* species, *Edgetheca* sp., *Crossbitheca* sp., *Tumulduria* sp., *Hyolithellus tenuis*, and *Barskovia hemisymmetrica*. Taxa mentioned above as characteristic of the *sunnaginicus* Zone in the eastern sections are represented by *Igorella hamata, Prooneotodus* sp., *Anabarites kelleri, Aldanella crassa, Spinulitheca rotunda*, and *Turcutheca rugata*. In the lower 3–4 m of the red-bed portion of the Variegated Formation, taxa known from the *sunnaginicus* Zone occur in association with *Fomitchella acinaciformis, Kotuitheca, Mellopegma* sp., *Conotheca mammilata, Anabarella plana*, and first *Nikatheca kengedeica*. The overlying 15-m-thick part of the Variegated Formation is poorly exposed. In this interval, rocks show a further diversity increase and yield forms known near the base of the

Variegated Formation in the middle course of the Aldan River. These are *Edgetheca aldanica, Coleolella* sp., *Camenella* sp., *Spinulitheca billingsi*, and *Tommotia admiranda*. They associate with *Allatheca cana* and with SSF of the *regularis* Zone: *Allatheca corrugata, Eonovitatus superbus, Burites distortus, Novitatus simplex*, and others. The overlying 4 m of the sequence yield, in addition to *Tommotia kozlowskii, T. plana*, and *Camenella garbowskae*, species of the *L. bella* Subzone: *Gracilitheca bayonet, Crestjachites figuratus, Obliquatheca bicostata*. Directly above this interval, first SSF taxa and trilobites of the Atdabanian Stage have been discovered (Val'kov, 1987; Missarzhevskii, 1989; Rozanov *et al.*, 1992). A complicated geological structure of this poorly exposed area does no afford to establish the correct stratigraphic position of these fossils in the section.

The Tommotian sections of transitional type can be traced from the Aldan River basin to the Lena River valley, at least from the Isit' Settlement to the Oi-Muran Settlement. The molluscan remains (Coreospiridae gen. et sp. indet.), hyolithids (Circothecidae gen. et sp. indet.), problematics (*Cambrotubulus* sp.), and algae (*Korilophyton inopinatum, Renalcis gelatinosus*) found in the upper 10 m of the Ust'-Yudoma Formation (*Yarusnoe raschlenenie...*, 1984) does not indicate that these layers belong to the Tommotian Stage, but not to the *antiqua* Zone. According to predominant gray and greenish gray coloration of limestones in the lower 7-m-thick interval of the Variegated Formation, it can be arbitrarily referred to the Sunnagin Member. The SSF assemblage of the *sunnaginicus* Zone from these limestones and basal red beds of the Variegated Formation is impoverished, as compared with that from the section exposed in middle courses of the Aldan River, and this stimulated Val'kov (1982) to refer it to the pre-Tommotian *A. cana* Zone. His viewpoint is unacceptable, however, because archaeocyathids of the Sunnagin Member and *Tommotia kozlowskii* were found 1–1.5 m above the base of the Variegated Formation in the borehole sections located opposite the Isit' and Sailyk settlements (Zhuravleva *et al.*, 1983). In this section, the lower boundary of the Tommotian Stage should likely be located with a hiatus at the base of the presumable Sunnagin Member.

All data on distribution of SSF assemblages in the Vendian–Cambrian boundary deposits of the southeastern Siberian platform lead to several definite conclusions. First, the numerous pre-Tommotian SSF zones, which were defined by Val'kov (1987) and Missarzhevskii (1989) in the region, are invalid. They cannot be used for correlation of sections in the transitional and eastern facies zones, and completely entangle positioning of boundary between the Tommotian and Nemakit-Daldynian stages there. Some zones with common names have different ages in the transitional and eastern facies zones. In the transitional facies zone, for instance, the Khayalakhan Stage after Val'kov, or the upper Manykaian Substage after Missarzhevskii, corresponds to Member IV of the Ust'-Yudoma Formation, or to the *antiqua* Zone of the Nemakit-Daldynian Stage, in which we identified up to 30 SSF taxa (Khomentovsky and Karlova, 1993). We do not propose here a more detailed subdivision of the *antiqua* Zone. Characteristic taxa of this zone from the Gonam River, Mt. Konus, and Dzhanda River sections are correlative with SSF assemblages from the basal part of the Koril Member of the Anabar region or with the *Anabarella plana* Zone of the North Siberian platform (Semikhatov and Serebryakov, 1983, Khomentovsky *et al.*, 1983; Khomentovsky and Karlova, 1993). This is supported by occurrence of *Anabarella plana* in Member IV of the Ust'-Yudoma Formation in the Dzhanda River section (Pel'man *et al.*, 1990).

The lower part of the Variegated Formation in sections exposed along the Selinde, Dzhanda, Aim, Maya, and Yudoma rivers of the eastern facies zone contains many taxa typical of the Ust'-Yudoma Formation. This fact and species of *Angustiochrea lata, Spinulitheca rotunda, Purella cristata, Aldanella crassa*, and *Anabarella plana* zones defined by Val'kov and Missarzhevskii in the northern facies zone, which occur at this level, were taken for a good reason to refer the lower part of the Variegated Formation to the pre-Tommotian interval. However, abundant guide forms of the *sunnaginicus* Zone and occurrence of archaeocyathids and SSF taxa of the *regularis* Zone in overlying rocks disprove this conclusion. Otherwise, we should admit that the *plana* Zone may occur below the marker Sunnagin Member and correspond to the pre-Tommotian *antiqua* Zone in the transitional facies zone, whereas in the eastern zone we should this and all other "pre-Tommotian" zones to overlie the latter and thus to be enclosed in the Tommotian Stage. To resolve this contradiction, Val'kov and Missarzhevskii limited the *regularis* Zone, contrary to factual data, leaving in it only the upper subzone (*L. bella*) and uniting the lower subzone with the *sunnaginicus* Zone into a single unit, a mishmash and nothing else. Missarzhevskii (1989) refers the *plana* Zone to the Tommotian Stage in the middle courses of the Aldan and Lena rivers, and to the Manykaian Stage in other areas of the Siberian platform. Val'kov (1987) draws the lower boundary of the Tommotian Stage in the middle of the *cana* Zone in one sections, but at its roof and even in the upper part of the overlying *Spinulitheca billingsi* Zone in others. It is interesting that they both interpret the time-range and position of the *A. plana* Zone in different ways. Missarzhevskii considers it as crowning the pre-Tommotian stage, whereas Val'kov correlates it with the *S. rotunda* Zone, the middle unit of the pre-Tommotian Khayalakhan Stage. Such discrepancy simply means that the so-called zones, defined in the eastern facies zone, characterize accidental co-occurrence of SSF taxa and cannot be accepted for biostratigraphic units.

It is evident that deposits of the early Tommotian time in the southeastern Siberian platform recorded mixing of a peculiar Siberian biota from the transitional

facies zone with the skeletal fauna of the eastern open basin. The eastward migration of Siberian taxa was distinctly lagged behind the moment of its first appearance. In this connection, different Siberian-type SSF taxa from the *sunnaginicus* Zone, the "determining" taxa included, spread in the Tommotian succession gradually. Many of them first appeared only together with some forms of the *regularis* and *lenaicus* zones, and even with those of the Atdabanian Stage. It is noteworthy that assemblages of the *sunnaginicus* Zone are everywhere more abundant and diverse than those of the *antiqua* Zone. This means that the lower Tommotian boundary marks the commencement of an extremely important stage in biota evolution.

It is remarkable as well that all, even most diverse assemblages of the *antiqua* Zone lack taxa of the *sunnaginicus* Zone. This fact is more important for the establishment of the lower boundary of the Tommotian Stage than the presence or absence of any "determining" taxon in the assemblage of the *sunnaginicus* Zone.

The peculiar composition of fossil assemblages of the Tommotian Stage in the eastern facies zone necessitates elaboration of a specific eastern zonal scale. As we could see, the above-mentioned zones suggested by Val'kov and Missarzhevskii are inappropriate for this purpose. More acceptable is a zonal scheme elaborated by Sysoev (1972) who proposed to subdivide the Tommotian Stage of the eastern facies zone into the following subunits (from the base upward): *Spinulitheca billingsi, Notabilites simplex,* and *Dorsojugatus sedecostatus* zones (the latter with subzones). He correlated more or less correctly these zones with biostratigraphic units of classical sections of the Tommotian Stage. In areas where the entire Tommotian section is characterized only by specific taxa of the eastern facies zone and shows a gradual appearance of only rare SSF species of the Siberian-type *sunnaginicus* Zone, we suggest to accept our *Fomitchella infundibuliformis* Zone corresponding to the whole Tommotian Stage (Khomentovsky and Karlova, 1992; Khomentovsky *et al.*, 1998). Smaller-scale units similar to those of the stratotype section can be outlined using an approach illustrated in Fig. 2. For this purpose, we bracketed the interval with most distinct lithological and paleontological boundaries, using in our case the basal levels of the Sunnagin Member and Atdabanian Stage, in the profile based on numerous sections of the Tommotian Stage. Further, taxa of a high correlative potential were designated by indices within this interval, and the first occurrence levels of youngest zonal assemblages were then extrapolated to sections, in which these assemblages can be expected.

THE NORTH SIBERIAN PLATFORM

In northern Siberia, one section of undoubtedly transitional type is exposed along the Sukharikha River at the Igarka Uplift (Figs. 1, 3). The upper 2 m of gray-colored carbonate rocks of the Sukharikha Formation and lower 1–2 m of the overlying Krasnyi Porog Formation lithologically similar to the Variegated Formation contain here archaeocyathids of the *sunnaginicus* Zone (Rozanov *et al.*, 1969). In addition to several transit taxa of the *antiqua* Zone, the SSF assemblage includes forms of the *sunnaginicus* Zone, such as *Tommotia kozlowskii, Camenella garbowskae, Edgetheca aldanica, Watsonella sibirica, Mellopegma indecora, Spinulitheca billingsi*, and others (Rozanov *et al.*, 1969; Missarzhevskii, 1989; Rowland *et al.*, 1998). Although these forms are typical of the *sunnaginicus* Zone, the underlying deposits are barren of fossils characteristic of the *antiqua* Zone, and the Sukharikha River section is inappropriate for substantiation of the lower boundary of the Tommotian Stage. Like in the Aldan River section, overlying beds enclose single *Lapworthella tortuosa* and, slightly higher, archaeocyathids and SSF taxa of two subzones of the *regularis* Zone (Rozanov *et al.*, 1969; Rowland *et al.*, 1998). Therefore, the idea of Missarzhevskii (1989) to unite the *sunnaginicus* Zone and lower subzone of the *regularis* Zone into a single unit of this area seems strange.

In boreholes K-203, K-204, and Eiik drilled in water divide areas between the Olenek, Muna, and Tyung rivers (Figs. 1, 3), i.e., near the boundary between the transitional and eastern facies zones (Khomentovsky *et al.*, 1998), beds with the SSF assemblage of the *P. antiqua* Zone (*P. antiqua, Anabarites signatus, Platisolenites* sp., and others) are overlain by a member bearing the same taxa and several forms of the *sunnaginicus* Zone of both the transitional (*Turcutheca crasseocochlia, Hyolithellus tenuis, H. grandis, Aldanella attleborensis, Barskovia hemisymmetrica, Securiconus costulatus*, fragments of undeterminatable archaeocyathids) and eastern (*Fomitchella acinaciformis, Aldanella crassa, Curtitheca korobovi*) types. Higher in the section, the number of taxa characteristic of the *sunnaginicus* Zone of both types increases, and *Aldanella rozanovi, Mellopegma indecora*, and *Nikatheca* sp. coexist here with *Gracilitheca* sp., *Eonovitatus* sp., and *Burites* sp. of the *regularis* Zone. Like in sections of the southeastern Siberian platform, the number of taxa characteristic the transitional zone sharply decreases in the *sunnaginicus* Zone of aforementioned borehole sections ("determining" taxa are missing) and the eastern SSF taxa appear instead of them. Many species of the *sunnaginicus* Zone first appear here only at the level of the *regularis* Zone (Khomentovsky *et al.*, 1998).

Worthy of particular attention in North Siberia is the northwestern part of the Anabar massif, where biostratigraphic profile of Vendian–Cambrian deposits extends from the Moiero River mouth to the Kotuikan River mouth area, and further northwestward to the Rassokha River (Figs. 1, 3). Data on sections along this profile have been used to substantiate and correlate the lower boundary of the Tommotian Stage in the southern and northern parts of the Siberian platform (Val'kov, 1987; Missarzhevskii, 1989; Kaufman *et al.*, 1996). At the Kotui River (upstream of the Moiero River), light-

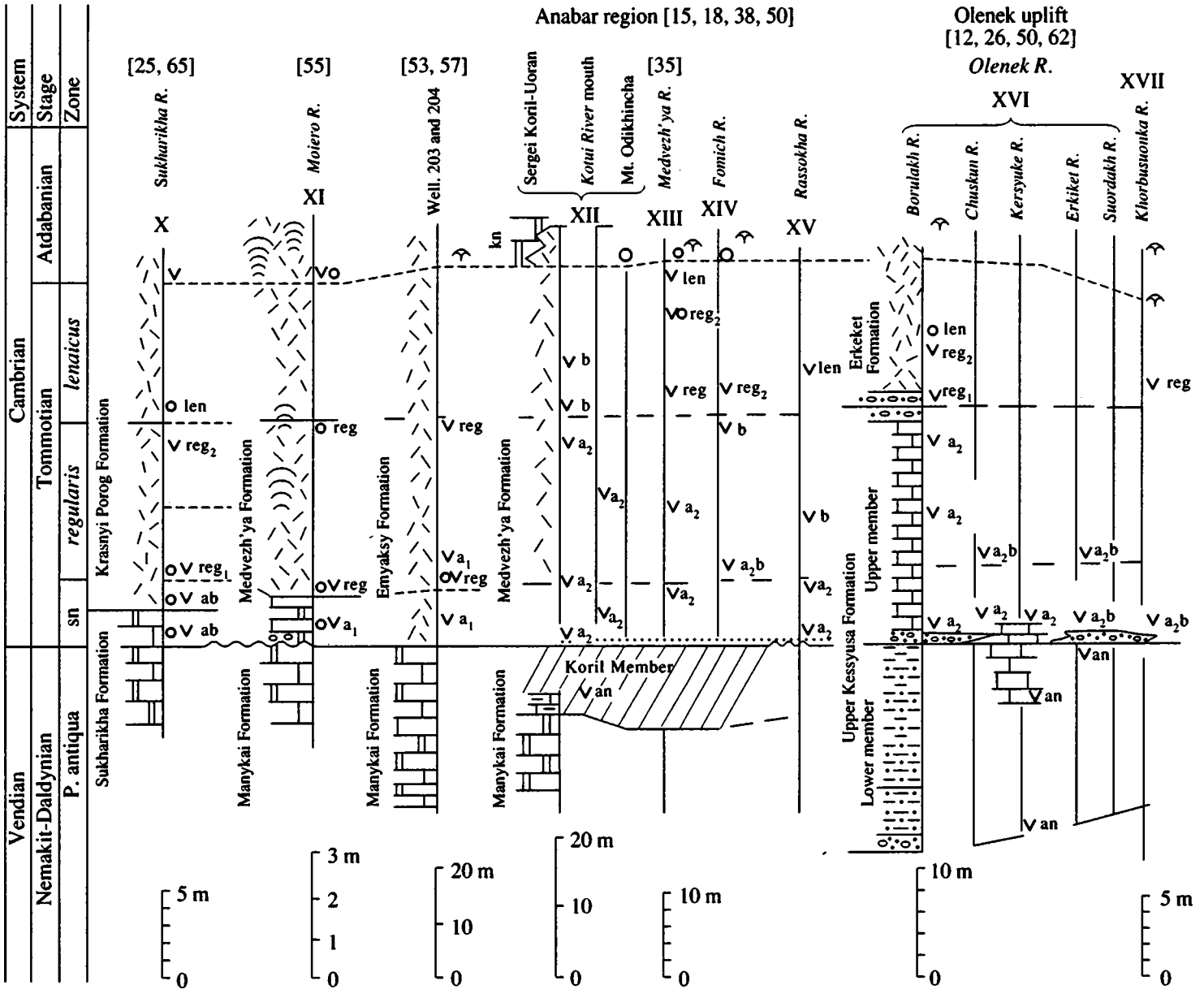

Fig. 3. Correlated sections and fossil assemblages of the northern Siberian platform (symbols as in Fig. 2). References: [15] Meshkova *et al.*, 1976, [18] Missarzhevskii, 1989, [25] Rozanov *et al.*, 1969, [26] Rozanov *et al.*, 1992, [35] Sundukov and Fedorov, 1986, [38] Fedorov and Shishkin, 1984, [50] Khomentovsky and Karlova, 1992, [53] Khomentovsky *et al.*, 1998, [55] Shishkin *et al.*, 1982, [57] Brasier and Surhov, 1998, [62] Khomentovsky and Karlova, 1993, [65] Rouland *et al.*, 1998.

colored carbonate rocks of the Staraya Rechka Formation are overlain, with a hiatus, by the terrigenous–carbonate Manykai Formation, which is overlain, in turn, by the varicolored Medvezh'ya Formation. The lower 1.2-m-thick member of the latter is composed of greenish gray dolomitic limestones with lenses of syndepositional conglomerates. Its layers bearing granular glauconite and abundant fossils resemble those of the Sunnagin Member in the Aldan River sections. Fossils from the member and overlying rocks, which are represented by archaeocyathids and SSF taxa of the *sunnaginicus* Zone (Shishkin *et al.*, 1982), emphasize the similarity. In addition to transit taxa *Purella antiqua, Cambrotubulus decurvatus*, and others, the SSF assemblage includes several species characteristic of coeval deposits in the Dvortsy and Ulakhan-Sulugur sections: *Bemella parula, Igorella monstrosa, Hyolithellus tenuis, Torellella lentiformis, Coleoloides trigeminatus, Conotheca mammilata.* The assemblage is remarkably less diverse as compared with that of the *sunnaginicus* Zone in middle courses of the Aldan River. First *Lapworthella tortuosa* and *Tommotia* forms unidentifiable at the species level appear here only at the top of layers under consideration. Beds immediately above them yielded archaeocyathids of the *regularis* Zone (*Nochoroicyathus mirabilis, Dictiocyathus translucidus, Robustocyathus robustus, Cambrocyathellus tschuranicus*) and abundant *Lapworthella tortuosa* and *Tommotia plana* (Shishkin *et al.*, 1982). Thus, the entire Medvezh'ya Formation, including its basal part, can be referred to the Tommotian Stage. Northeast of the Moiero River mouth, a hiatus becomes evident at the base of the Manykai Formation, although in the upper part of the Rassokha River this formation still rests conformably upon the Koril Member of dolomitic limestone that is 12–14 m thick and traceable from here to the Kotui River. This member disappears from the section only in lower courses of the Rassokha River (Missarzhevskii, 1989). The aforesaid repudiates his statement that a noticeable hiatus can be traced at the Medvezh'ya Formation base in the area between the Kotui River and middle courses of the Rassokha River. Also erroneous is an attempt (Missarzhevskii, 1989) to argue for a progressive northeastward pinch-out of the lower parts of the Medvezh'ya Formation corresponding to the interval of *Purella cristata–Anabarella plana* zones that is supposedly related to this hiatus. Lithological data used to substantiate this viewpoint were incorrectly interpreted. In the area between the Sergei Koril-Uoran Rapid of the Kotui River and the Chomn-Yuryakh Creek, the lower 10- to15-m-thick interval of the Medvezh'ya Formation is composed of alternating red, greenish gray, and gray clayey limestones. The bed succession in this interval sharply changes along a distance of few dozens of meters. Our data support conclusion of Fedorov and Shishkin (1984) that the variegated member does not wedge out between the rapid and mouths areas of the Kotuikan and Kugda rivers; to the contrary, its thickness increases here.

In the greater part of the Anabar region, where the marker Koril Member is easily identifiable, its paleontological characteristic principally differs from that of the basal part of the Medvezh'ya Formation. The SSF assemblage of the Koril Member includes about 20 taxa, excluding transit forms of the *A. trisulcaus* Zone (Khomentovsky and Karlova, 1992, 1993). This assemblage is similar to that of the *P. antiqua* Zone of the southeastern Siberian platform. Coexistence of *Purella antiqua, P. cristata, Ladatheca dorsocava, Anabarella plana, Latouchella korobkovi, Angustiochrea lata*, and others in this assemblage (Val'kov, 1975, 1987; Missarzhevskii, 1989; Fedorov and Shishkin, 1984; Khomentovsky and Karlova, 1992) indicate that it includes the guide forms of the *A. lata, P. cristata*, and *A. plana* zones defined by Val'kov and Missarzhevskii. Missarzhevskii (1989) included into the *cristata* Zone of the Sergei Koril-Uoran section at the Kotui River also the lower 9 m of the Medvezh'ya Formation, but this interval contains, according to other data (Meshkova *et al.*, 1976; Fedorov and Shishkin, 1984), such species as *Aldanella crassa, A. utchurica, A. rozanovi, A. attleborensis*, and *Hyolithellus tenuis*, in addition to other transit forms. Rocks of this member, which are exposed along the Ary-Mas-Yuryakh River, in the Kotuikan River mouth area, and along the creeks flowing from Mt. Odikhincha into the Kotui River, also yield *Coleolus trigonus, Igorella ungulata, Turcutheca crasseocochlia, Bemella jacutica, B. septata, Torellella* sp., and *Exilitheca* sp. The overlying beds (5–6 m) contain Tommotidae, *Conotheca mammilata, Sunnaginia* sp., *Exilitheca multa, Lapworthella* sp., *Torellella lentiformis*, and *Igorella hamata*, all unknown in pre-Tommotian deposits. In the interval of 16–35 m above the base of the Medvezh'ya Formation, which is referred to the pre-Tommotian *A. plana* Zone in the Sergei Koril-Uoran section (Missarzhevskii, 1989), the SSF assemblage shows no substantial compositional changes. In beds occurring 5 m higher, Missarzhevskii registered the first occurrence of *Tommotia kozlowskii, Camenella garbowskae, Spinulitheca billingsi*, and *Allatheca concinna*; based on these finds, he placed the lower Tommotian boundary at the base of this member. Other researchers (Val'kov, 1987; Meshkova *et al.*, 1976) failed to find the mentioned taxa at this level, and they seem to be scarce here. Last SSF were found in this section at the level of 55 m above the base of the Medvezh'ya Formation. The corresponding bed represents a peculiar marker and yields *Tchuranitheca simplex, Curtitheca korobovi, Allatheca concinna*, and *Hyolithellus tenuis.* Further to the north, the level of 55–60 m above the base of the Medvezh'ya Formation that is exposed along the Kugda-Yuryakh Creek and on the western slope of adjacent Mt. Odikhincha yielded archaeocyathids of the Atdabanian Stage (Meshkova *et al.*, 1976). In the Medvezh'ya and Fomich river heads, the lower boundary of the Atdabanian Stage is located, judging from the appearance of archaeocyathids of the *Leptosocyathus plyseptus* Zone and tri-

lobites *Profallotaspsis* (Repina and Luchinina, 1981), at the level of 45–48 m above the base of the Medvezh'ya Formation, i.e., somewhat lower than in Odichincha section. This can be explained by thickness reduction rather than by a hiatus at the base of the formation, because its lower 7–8 m contain the SSF assemblage identical to that in the Sergei Koril-Uoran section and consisting of *Aldanella rozanovi, A. crassa, Bemella jacutica, Hyolithellus tenuis*, and other forms (Fedorov and Shishkin, 1984). Further northeastward at the Rassokha River, the Tommotian Stage is even less thick, but the basal 10 m of sequence analogous to the Medvezh'ya Formation correspond, judging from finds of *Aldanella* sp. and *Hyolithellus tenuis* (Meshkova *et al.*, 1976), to the stratigraphic level identical to that in the Kotui River section despite the fact that they rest directly upon the Mukun Formation. In the Medvezh'ya River section, the assemblage of "determining" taxa with tommotids and *Edgetheca aldanica* was not found at all, but the section demonstrates the succession of SSF assemblages from levels of 31, 41, 46, and 48–50 m, above the base of the Medvezh'ya Formation, which characterize respectively the *regularis* (presence of *Lapworthella tortuosa* and *Allatheca corrugata*), *L. bella*, and *D. lenaicus* zones crowned by the Atdabanian Stage (Sundukov and Fedorov, 1986). At the Fomich River, different *Tommotia* species and *Camenella garbowskae* characterize the interval of 10–25 m above the Medvezh'ya Formation base; higher units are the 5-m-thick *L. bella* Zone and the *D. lenaicus* Zone above it (Fedorov and Shishkin, 1984; Missarzhevskii, 1989).

Thus, in the northwestern Anabar region, the lower 9–15 m of the Medvezh'ya Formation are characterized by many SSF species, which are known in the *A. sunnaginicus* Zone of the stratotype area, being barren of the so-called "determining" taxa. The latter (different *Tommotia* species, *Edgetheca aldanica, Allatheca concinna*, and others) sporadically occur only in the uppermost Tommotian part of the Medvezh'ya Formation on the west. The abundance of "determining" taxa and their distribution range increase in the northeast part of the biostratigraphic profile under consideration. The remarkable feature of beds bearing fossils of the *D. regularis* and *D. lenaicus* zones is their thickness that is universally insignificant, as compared with that of beds bearing only fossils typical of the *sunnaginicus* Zone stratotype. The corresponding thickness ratio locally equals 1/10. To explain this proportion, Missarzhevskii (1989) considered the *sunnaginicus* Zone and *tortuosa* Subzone as a single unit, and referred the lower part of the Medvezh'ya Formation to the pre-Tommotian *Aldanella crassa* and *Anabarella plana* zones. In our opinion, the Anabar region demonstrates compositional changes in SSF assemblages similar to those observed in the southeastern Siberian platform in direction from the transitional to eastern facies zone. As is shown, assemblages of transitional type are characteristic only of sections exposed near the Moiero River mouth. Species *Aldanella crassa, Fomitchella infundibuliformis, F. acinaciformis, Turcutheca cotujensis, Nikatheca kengedeica, Kotuitheca curta, Spinulitheca rotunda*, and many others, which appear already in basal layers of the Medvezh'ya Formation and are unknown below the Sunnagin level, suggest that Tommotian Stage in other areas of the Anabar region characterizes the eastern facies zone of that time.

As a result, a paradoxical situation is on hand. As shown, the base of the Koril Member near the mouth of the Kotuikan River contains the assemblage of the *plana* Zone (Val'kov, 1975; Fedorov and Shishkin, 1984; Khomentovsky and Karlova, 1992). On the other hand, Missarzhevskii (1989) defines his older *A. crassa* and *P. cristata* zones above this member of the Medvezh'ya Formation, but below the *plana* Zone. In such a case, the *plana* Zone appears defined at two principally different stratigraphic levels. Moreover, the real stratigraphic distribution of SSF taxa does not accord with zonal schemes suggested by Val'kov and Missarzhevskii. The same disagreement was demonstrated above for the southeastern part of the platform, and we cannot accept their zonations. Our statement that the Medvezh'ya Formation begins with the *sunnaginicus* Zone is evident from occurrence of guide archaeocyathids and SSF taxa already in basal layers of the formation in the Moiero section (see above). A sharp decrease in thickness of upper SSF-bearing Tommotian beds can be explained, like in the southeastern sections, by the delayed migration of taxa from the transitional facies zone into the eastern one (Khomentovsky and Karlova, 1992, 1994). The *tortuosa* Subzone of the *regularis* Zone has not been defined, and the *bella* Subzone has been united with the *lenaicus* Zone (Val'kov, 1987; Missarzhevskii, 1989), because these taxa turned out to be missing in continuous sections. The aforesaid implies that only further investigation of SSF assemblages may solve the problem of the Tommotian Stage biozonation in both the eastern facies zone and northern areas.

The conclusion that the lower part of the Medvezh'ya Formation corresponds to the *sunnaginicus* Zone despite the scarcity of tommotids and some other SSF taxa at this level seems to be valid in the light of data from the Olenek uplift (Figs. 1, 3). Missarzhevskii (1989) failed to find, despite the SSF abundance, the "determining" taxa (unknown below the *sunnaginicus* Zone) in the upper member of the Upper Kessyusa Subformation of the Olenek River valley between mouth areas of the Borulakh and Kersyuke rivers. Accordingly, he concluded that relevant deposits are eroded here, and that beds with SSF assemblages of the pre-Tommotian *crassa* and *plana* zones are overlain by the Ereket Formation. The lower interval of the latter (3–5 m) yielded *Allatheca corrugata, Burites distortus*, and other taxa typical of the *regularis* Zone, whereas forms characteristic of the *bella* and *lenaicus* zones were discovered in layers immediately above it (Missarzhevskii, 1989; Rozanov *et al.*, 1992). The level of

20 m above the base of the formation yielded trilobites of the Atdabanian Stage. However, in the section near the Ereket River mouth, slightly downstream of the Olenek River, Missarzhevskii found, along within the SSF assemblage referred to the *plana* Zone, species *Tommotia kozlowskii, Watsonella* sp., and Allathecidae in the upper member of the Upper Kessyusa Subformation and concluded that the basal part of the Tommotian stage is preserved in this section. Tracing marker beds in several intermediate sections, we established that below the Eretek Formation the upper member of the Upper Kessyusa Subformation has whole range in all sections of the Olenek River basin between the Borulakh and Ereket creeks and downstream (Khomentovsky and Karlova, 1992). Even in sections located far upstream, basal layers of this member contain, in addition to taxa of the so-called *crassa* and *plana* zones (Missarzhevskii, 1989), *Barskovia hemisymmetrica, Allatheca concinna, Aldanella rozanovi, Conotheca mammilata*, and *Watsonella sibirica*, which are considered to be typical of the Tommotian Stage in the Aldan River sections. All sections also enclose many taxa characteristic of the *sunnaginicus* Zone in the eastern facies zone (Missarzhevskii, 1989; Khomentovsky and Karlova, 1992; Rozanov *et al.*, 1992). At the Khorbusuonka River, the same stratigraphic level yielded in addition rare *Edgetheca aldanica* and *Tommotia admiranda*. Many tommotids and other "determining" taxa appear in abundance only in basal layer of the Eretek Formation that contains, as is mentioned, guide forms of the *regularis* Zone. Abundant trilobite remains in preparations from the Khorbusuonka River section are discovered 6 m above the base of Eretek Formation, and genus *Profallotaspis* first appears 5 m higher (Rozanov *et al.*, 1992).

Summing up the data on biostratigraphy of the Vendian–Cambrian deposits of the northern Siberian platform, we should emphasize that the most distinct biostratigraphic boundary in this region is the base of the *sunnaginicus* Zone, or the base of the Tommotian Stage. While the *antiqua* Zone of the Anabar region and Olenek uplift marks the appearance of 30 typical SSF taxa, the *sunnaginicus* Zone hosts 75 characteristic forms (Khomentovsky and Karlova, 1992, 1993). Proportions are approximately the same as in transitional and eastern facies zones in the southern areas. Many SSF taxa of the *antiqua* and *sunnaginicus* zones are common for the southern and northern parts of the Siberian platform. All this points to a principal significance and synchronism of the boundary separating these zones.

The other quite distinct boundary in the northern part of the platform separates the Tommotian and Atdabanian stages. Extremely important for its substantiation are occurrence levels of archaeocyathids and trilobites (Fig. 3). It is more difficult to subdivide the Tommotian Stage, because many forms characteristic of the *sunnaginicus* Zone transit into younger strata, and rare "determining" taxa of the *sunnaginicus, regularis*, and *lenaicus* zones appear in sections of the eastern facies zone with a great delay. Taxa of the first zone are discovered to occur frequently together with characteristic forms of the upper Tommotian beds, which are usually significantly less thick than lower ones. Often, it is impossible to discriminate the *lenaicus* and *regularis* zones based on SSF assemblages or to subdivide the latter into the *tortuosa* and *bella* subzones. Correlation of reference sections in the northern part of the platform is also not an easy task because of a highly variable thickness of the Tommotian Stage in different structural-facies zones. That is why we had to vary the column scale in our correlation scheme of the Tommotian Stage (Fig. 3) in order to make them comparable in dimension for synchronous parts of sections between basal planes of the *sunnaginicus* Zone and Atdabanian Stage. Indicating fossils in this scheme and taking into consideration the first appearance level of youngest zonal forms, we managed to discriminate, in general, intervals in the northern sections of the Tommotian Stage, which correspond to *sunnaginicus, regularis*, and *lenaicus* zones. This is the only way to define arbitrarily analogues of biostratigraphic units of the Tommotian Stage in the transitional zone of the southern Siberian platform.

As is shown below, chemostratigraphic data confirm our conclusion on the position of the lower boundary of the Tommotian Stage in the Anabar region. Thus, all available factual materials indicate that correlation of Tommotian sections of the northern and southern areas of the Siberian platform proposed here is more substantiated that suggested by other researchers (Val'kov, 1987; Missarzhevskii, 1989; Kaufman *et al.*, 1996).

ON THE LOWER BOUNDARY STRATOTYPE OF THE TOMMOTIAN STAGE

The transitional-type sections are most appropriate for determination of the lower boundary of the Cambrian System, because basal layers of the latter contain, in addition to diverse SSF assemblages, archaeocyathids representing the orthostratigraphic group of the Tommotian Stage. Therefore, the contact between the Ust'-Yudoma and Variegated formations that is well exposed at the bank of the Aldan River near the Ulakhan-Sulugur Creek mouth is very suitable in this respect (Rozanov *et al.*, 1969). However, investigations in this area provoked a discussion concerning the position of boundary between the Tommotian and Nemakit-Daldynian stages. Most debatable is origin of glauconite sandstones that are exposed 1.5 m below the roof of the Ust'-Yudoma Formation in the Ulakhan-Sulugur section and contain archaeocyathid remains in association with a representative assemblage of fossils characteristic of the *sunnaginicus* Zone, such as *Aldanella rozanovi, Barskovia hemisymmetrica, Edgetheca aldanica, Conotheca mammilata, Sunnaginia imbricata*, and others. Some researchers believe that these sandstones and dolomites of Bed 8 enclosing them are

syngenetic (Rozanov *et al.*, 1969; Fedorov *et al.*, 1992; Zhuravlev, 1998), while others identify them, following J.D. Aitkin, with sedimentary dikes, which transported fossils of the Variegated Formation together with clastic material into karst hollows in the Ust'-Yudoma Formation (Mens, 1975; Avdeeva *et al.*, 1983; Khomentovsky *et al.*, 1990; Khomentovsky and Karlova, 1993).

We established (Khomentovsky *et al.*, 1990; Khomentovsky and Karlova, 1993) that largest bodies of glauconite–dolomitic sandstones, which are similar in lithology to basal layers of the Variegated Formation, originate from Bed 7, but not from Bed 8 marked in exposed cliffs (*Putevoditel' ekskursii...*, 1973), being traceable only for a distance of 7 m. Sandstone bodies of capricious outlines are 5–10 cm thick, sometimes isometric and angular, oriented obliquely relative to bedding planes. They yield the majority of SSF taxa of the so-called Bed 8. In true Bed 8, lentils of granular glauconite are rare and thin, showing prevalence of dispersed glauconite; along the strike, they become more extended. Crosscutting bodies with glauconite and their horizontal tongues occur throughout the section up to the base of the Variegated Formation (Avdeeva *et al.*, 1983; Khomentovsky *et al.*, 1990; Khomentovsky and Karlova, 1993). These data, as well as poorer preservation and roundness value of SSF remains that is higher in glauconite-bearing dolomitic sandstones of Bed 7 than at the base of the Variegated Formation, suggest that fossils were transported together with clastic material into karst hollows of Bed 7 via a complicated system of sedimentary dikes. Abundant empty hollows in Bed 7 whose outlines are similar to those of glauconite sandstone bodies support this inference (Khomentovsky *et al.*, 1990). Some difference in composition and iridium content of glauconites from sandstones of Bed 7 and basal layers of the Variegated Formation can be a consequence of grains differentiation in accord with their specific weight. Therefore, this and other features (Fedorov *et al.*, 1992) do not contradict the accepted concept. Another fact in favor of our concept is that fossils other than *Hyiolithellus, Chancelloria*, and *Markuelia* species typical of the *P. antiqua* Zone have never been found, despite the very attentive search, in dolomites enclosing bodies of glauconite sandstones with abundant Tommotian SSF remains (Avdeeva *et al.*, 1983). It is also noteworthy that a hiatus at the base of the Variegated Formation in the Ulakhan-Sulugur section eliminated upper 5 m of the Ust'-Yudoma Formation, which are preserved in the Dvortsy exposure (Avdeeva *et al.*, 1983; Semikhatov and Serebryakov, 1983; Khomentovsky and Karlova, 1993) and, what is even more important, SSF taxa of the *sunnaginicus* Zone are absent below this hiatus.

Therefore, more valuable for determination of the lower Tommotian boundary is information on paleontologic characteristics of boundary layers between the Ust'-Yudoma and Variegated formations in the Dvortsy section, where a hiatus between these units is either completely missing, or less significant than in the Ulakhan-Sulugur section (Fedorov *et al.*, 1992; Khomentovsky and Karlova, 1992). Fedorov *et al.* (1992) discovered *Aldanella rozanovi, Bemella costata, Barskovia hemisymmetrica, Hyolithellus tenuis, Torellella curvae, Conotheca mammilata*, and other taxa of the *sunnaginicus* Zone in association with indeterminable archaeocyathid remains from lentils of glauconite–dolomite sandstone situated 0.2–0.3 m below the base of the Variegated Formation. We believed that this SSF assemblage is similar in origin to that of Bed 7 (Khomentovsky *et al.*, 1990; Khomentovsky and Karlova, 1992). However, lower beds of the Sunnagin Member in the middle of the Dvortsy exposure are composed of light dolomitic limestones with thin lenses of glauconite sandstones similar to those, in which Fedorov and his colleagues detected SSF remains. Downward, they grade into the 1-m-thick member of pure light limestones and dolomites of the Ust'-Yudoma Formation that yields *Bemella jacutica, Halkieria sacciformis, Torellella curvae, Hyolithellus tenuis, H. vladimirovae, Turcutheca* sp., *Conotheca* sp., *Tiksitheca licis, Protospongia, Chancelloria*, and *Markuelia*. The number of Tommotian forms is lower here than the younger layer described by Fedorov with colleagues. We identified additional Tommotian species *Edgetheca aldanica, Isitiella* sp., and *Torellella lentiformis* in lowest lenses of glauconite sandstone of the Sunnagin Member. A similar assemblage is characteristic of lowermost beds of the Variegated Formation, the most complete sections of which is exposed the Vas'kina Channel and Byuktelekh sections, where it is composed predominantly of pure light limestones. In addition to the Tommotian forms mentioned above, these rocks yield *Aldanella utchurica, Gonamella rostrata, Turcutheca crasseocochlia, Tumulduria* sp., *Sunnaginia* sp., and *Coleolella billingsi*, in association with brachiopods *Aldanothreta* sp. and archaeocyathid fragments. Only 1.5–2.0 m above its base, the Variegated Formation contains here the diverse tommotids and other "determining" taxa. Their assemblage is in general similar to that from basal layers of this formation in the Ulakhan-Sulugur section.

Thus, below the condensed Sunnagin Member of the Ulakhan-Sulugur section, where it is separated from Ust'-Yudoma deposits by a distinct hiatus, the Variegated Formation of the Dvortsy section is crowned by strata about 1.5–2 m thick. According to lithological–stratigraphic and paleontological data, the latter span the range of hiatus established in the Ulakhan-Sulugur section, forming a gradual transition between the Variegated and Ust'-Yudoma formations. Nevertheless, such a rapid enrichment of section in diverse SSF is not a result of explosion-type evolution of skeletal fauna near the indicated boundary. Abundance of SSF can be explained by the fact that the almost barren Ust'-Yudoma Formation of the western facies zone is overlain here by the Variegated Formation that accumulated already in the transitional zone.

Taking into consideration the 5-m-thick eroded interval at the base of the Variegated Formation in the

Ulakhan-Sulugur section, one may conclude that bodies of glauconite sandstones of Bed 7 are located 5–6 m, but not 1.5 m below the formation base. This contradiction to data on SSF distribution in the Vendian–Cambrian interval of the Dvortsy section can be regarded as additional argument against possibility to place the lower Tommotian boundary at the base of that bed. Consequently, only the base of the Variegated Formation in the Dvortsy section, where the Tommotian Stage has been defined, can represent GSSP of this boundary in sections of the Aldan River middle courses.

Inasmuch as the hiatus at the base of the Variegated Formation in the excellently exposed and easily accessible Ulakhan-Sulugur section is insignificant, the later can also be considered as a candidate for the lower Cambrian boundary stratotype. In this case, however, one should, with due regard for data on the Dvortsy section, draw the boundary at the base of the Variegated Formation, but not at the base of Bed 7 of the Ust'-Yudoma Formation.

The classical Aldan sections, however, are inappropriate for the lower Tommotian boundary stratotype, because their weak point is a very poor paleontological characteristic of deposits underlying the *antiqua* Zone of the Nemakit-Daldynian Stage. The representative fossil assemblage of this zone is found only near its base in sections of the height with altitude of 1291 m at the Gonam River and at the Mt. Konus site. Sufficiently diverse SSF assemblages of the closely related *antiqua* and *sunnaginicus* zones are recorded in the Dzhanda River section (see above). However, in the assemblage of the *sunnaginicus* Zone, prevalent SSF taxa are those of the eastern facies zone, and the Dzhanda River section can be considered only as a parastratotype of the lower Tommotian boundary in the eastern facies zone of the southeastern Siberian platform. A much better candidate for the stratotype section is located in this zone between the Kersyuke and Kere-Yuryakh tributaries of the Olenek River, where successive and representative assemblages of the *antiqua* and *sunnaginicus* zones are detected in a continuous section (Khomentovsky and Karlova, 1992). In the section near the mouth of the Kotuikan River, the western Anabar region, both biostratigraphic units yield poor fossil assemblages, and position of the lower Tommotian boundary is highly debatable here. An important argument in favor of this boundary position between the Manykai and Medvezh'ya formations is a negative $\delta^{13}C$ peak recorded at this level (Khomentovsky, 1996; Khomentovsky *et al.*, 1998).

CONCLUSION

The oldest small shelly Metazoa represent a main transit group of fauna, evolution of which is important for correct determination of boundary between the Nemakit-Daldynian and Tommotian stages. Appearance of early Tommotian archaeocyathids and brachiopods at this level is also extremely important in this aspect, although is not a simple task to select any boundary stratotype based on the first occurrence levels of some representatives of skeletal fauna (Cowie and Glaessner, 1975). At the same time, is more difficult to discriminate Vendian–Cambrian biostratigraphic units and to substantiate their boundaries based on distribution of ancient SSF assemblages than to operate with evolutionary patterns of younger macrofaunas. The remarkable feature of almost all ancient SSF is that they first appear as single specimens, which become significantly more abundant later, sometimes even in a younger zone. Moreover, even insignificant lateral changes in habitats can result in complete disappearance of some SSF taxa or entire assemblages (Khomentovsky and Karlova, 1992, 1994). Therefore, one should be very careful by correlation of local sections with any standard using individual, even guide taxa. It is necessary first to sample SSF remains in all beds of many individual sections under examination and to compare their individual assemblages in order to establish real age relationships between them. Only in this case, correlation between local and standard sections can be adequate. We also should use a particular approach to substantiate correlation of regional biostratigraphic units in transitional areas between different sedimentation settings. The aforementioned multivariant combinations of taxa, characterizing coeval strata in different exposures of a single facies zone, become extremely entangled, when these taxa turn out to mixed with representatives of another facies zone. Any situation of this kind may result in recognition of some artificial zonal units. Therefore, it is very important to select the most substantiated standard and trace its units over a maximum possible distance The Vendian–Cambrian SSF biozonation established in the transitional facies zone of Siberia is undoubtedly a standard of this kind. The complex *antiqua* Zone of the Nemakit-Daldynian Stage, as well as *sunnaginicus* (*tortuosa* and *bella* subzones) and *lenaicus* zones of the Tommotian Stage are distinguished here based on archaeocyathid, brachiopod, and SSF assemblages mutually complementing and controlling each other. In the type sections of the eastern facies zone, archaeocyathids are absent at this stratigraphic level, and eastern SSF taxa first appearing at the base of the *sunnaginicus* Zone have wider stratigraphic ranges. Of an extreme importance here are even single finds of archaeocyathids and SSF remains characteristic of the transitional zone. Based on their first occurrence levels, we can prolong, with some reservations, units of the transitional zones in those areas, where only SSF taxa of the eastern zone are detectable (Figs. 2, 3). One should use, therewith, different lithological and stratigraphic markers as widely as possible. Owing to almost continuous distribution of Vendian–Cambrian deposits in the southeastern Siberian platform and in different structural–facies zones of its northern part, markers such as the Sunnagin Member of the Aldan massif of the Koril Member of the Anabar massif are of a great significance for the correct corre-

lation. We believe that SSF assemblages of Vendian–Cambrian biostratigraphic units distinguished in Siberia are most representative, and that synthesis of data available for this vast territory is demonstrative with respect to practical usage of this fossil group by solution of stratigraphic problems in other regions.

The approach discussed above was ignored by Landing (1994, 1996) who correlated, based on SSF assemblages, the Vendian–Cambrian biostratigraphic units of Siberia with those of the so-called Avalon Insular continent. In a succession (from the top downward) of the *Camenella baltica, Sunnaginia imbricata, Watsonella crosbyi (Aldanella attleborensis)*, and *Ladatheca cylindrica* zones defined by Landing, even the uppermost one is declared to be of the pre-Tommotian age in its larger part. However, *Gracilitheca* forms mentioned amid SSF remains of the zone appears only in the uppermost part of the Tommotian Stage, whereas *Rhombocorniculum cancellatum* represents a zonal form of the upper part of the Atdabanian Stage (Missarzhevskii, 1989). In addition, the overlying Brigus Formation contains trilobites of the Botomian Stage. The older *Sunnaginia imbricata* Zone cannot also be considered as the pre-Tommotian unit because its index species appears in Siberia only in the Tommotian Stage like the accompanying *Torellella laevigata, Lapworthella*, and *Fomitchella* (see above). There are no grounds as well to place the *Watsonella crosbyi* Zone below the Tommotian. Genus *Watsonella* (=*Heraultipegma*) is unknown in the *antiqua* Zone, whereas associated *Allatheca degeeri* and *Lapworthella* species appear in Siberia and China usually above the base of the Tommotian Stage. Even the oldest zone of the Avalonian succession cannot be confidently considered as the pre-Tommotian subdivision, because the *Ladatheca* genus characterizes mainly the Tommotian Stage in Siberia (Rozanov *et al.*, 1997).

In China, the lower boundary of the Cambrian System at the B point of the Meishucun section (Xing Yusheng *et al.*, 1991) is well substantiated based on the changes in SSF assembalges, as we believe, and corresponds to the base of the Tommotian Stage (Khomentovsky and Karlova, 1993). Simultaneously, it seems incorrect to term the underlying deposits as the *Anabarites–Protohertzina* Zone, because one gets impression that it corresponds to the *A. trisulcatus* Zone of Siberia. The greater part of this unit belongs undoubtedly to the *P. antiqua* Zone, and only the basal part of the Xiaowaitoushan Member and its analogues can be correlated with the *trisulcatus* Zone.

Despite the SSF abundance in the Soltaneih Formation of Iran (Hamdi *et al.*, 1989), its correlation with the Siberian standard is ambiguous. According to occurrence of *Hyolithellus* and Monoplacophora forms, its lowermost part (185 m) can be identified with the *antiqua* Zone. The uppermost layers in this interval of the section contain characteristic and abundant species *Purella* sp., *Tiksitheca licis, Ladatheca* sp., and *Maikhanella multa*. The base of the overlying Upper Shale Member (80 m thick) is marked by appearance of SSF forms characteristic of the Tommotian Stage (Allathecidae, *Aldanella* sp.), but strata located 20 m above the base of the member yield trilobites *Eoredlichina* and *Wutingaspis*, which are most likely in situ, because they are closely connected with the phosphorite-bearing horizon that represents an excellent marker. It is also noteworthy that *Latouchella korobkovi*, the characteristic taxon of the *antiqua* Zone, first appears in the Iranian section almost simultaneously with the Atdabanian *Pelagiella* forms. Thus, the situation here resembles that observed in some sections of the eastern facies zone of Siberia.

Recently obtained chemostratigraphic data confirm our correlation scheme for Vendian–Cambrian sections of the transitional and eastern facies zones. Moreover, owing to these data, biostratigraphic units characterizing this interval in the transitional zone can also be defined confidently in the reference section of the western zone near the Angara River mouth (Khomentovsky *et al.*, 1998a). Formerly, such a correlation was based on historical–geological data only. As one can see in Fig. 4, $\delta^{13}C$ variations within stratigraphic interval under discussion are synphase in all three facies zones. All the variation curves demonstrate a significant negative excursion (A) near the top of the Tommotian Stage. Less distinct is a positive excursion near the boundary between the *lenaicus* and *regularis* zones. Sharper negative peaks are characteristic of the middle part of the *regularis* Zone (B) and of the boundary beds between the *sunnaginicus* and *antiqua* zones (C). All curves also demonstrate some positive peaks near the boundary between the *regularis* and *sunnaginicus* zones and within the *antiqua* Zone (D).

An erroneous statement that the lower boundary of the Tommotian Stage corresponds to the upper part of the Medvezh'ya Formation (Kaufman *et al.*, 1996) resulted from the incorrect identification of $\delta^{13}C$ peak C recorded in the Kotuikan River section mouth with peak B of Dvortsy section (Fig. 4). This example demonstrates that correlations based on chemostratigraphic data should be supported usually by additional information. Paleontological and chemostratigraphic data (Khomentovsky *et al.*, 1998; Brasier and Surhov, 1998) made it possible to correlate with confidence the Vendian–Cambrian interval of core sections recovered by wells K-203 and K-204 in the water divide between the Olenek–Muna and Aldan river basins (Fig. 4).

Isotopic dates characterizing the lower Tommotian boundary are significantly less certain. Until recently, numerous determinations of K–Ar glauconite ages were used for this purpose; all of them fall into the interval between 540 and 600 Ma. When U–Pb ages of 536 to 531 Ma were obtained for zircons from volcanic pebbles sampled near the roof of the Lower Tyuser Subformation (Bowring *et al.*, 1993), these dates were declared invalid, however (Semikhatov, 2000). Age of

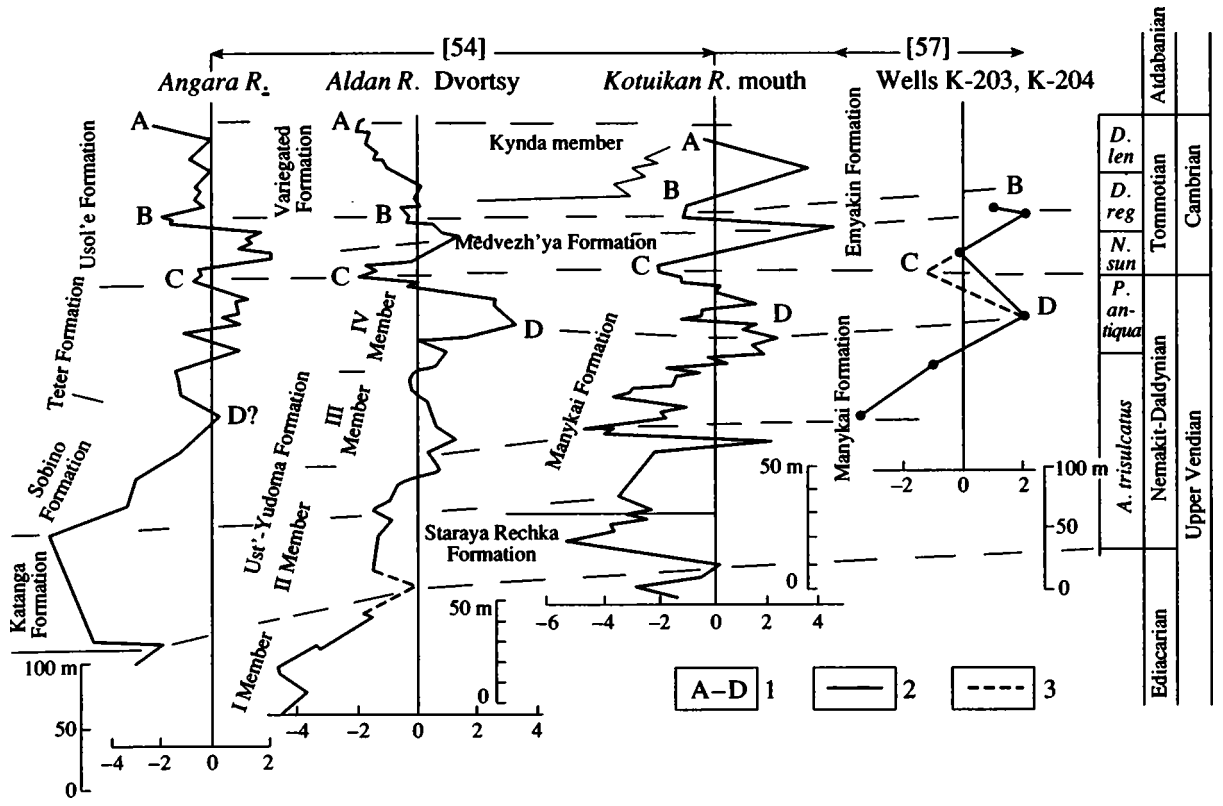

Fig. 4. Correlation of $\delta^{13}C$ excursions with stratigraphic units of reference Vendian–Cambrian sections in the Siberian platform: (1) correlated $\delta^{13}C$ peaks; (2) lithostratigraphic and (3) biostratigraphic boundaries. References: [54] Khomentovsky *et al.*, 1998a, [57] Brasier and Surhov, 1998, [61] Kaufman *et al.*, 1996.

the middle Tyuser Subformation was under debates during many years. Some researchers believed that the lower part of the subformation corresponds, according to SSF and archaeocyathid evidence, to the *regularis* Zone (Repina *et al.*, 1974). Others who discovered the complete sets of microfossils from the Talsin and Vergal' horizons considered them to be the Atdabanian in age and, moreover, correlated the Tommotian Stage with the *Holmia* Zone (Vidal *et al.*, 1993). Repina with colleagues noted, however, that in the Kharaulakh section, trilobites from second zone of the Atdabanian Stage (*Fallotaspis* Zone) occur in strata immediately overlying the *regularis* Zone. Their data and fact that archaeocyathids (*Dictyocyathus translucidus*) and all SSF taxa transit from the level in question into the Atdabanian Stage eliminate the contradiction. The matter is that the SSF taxa of the *regularis* Zone appeare in the Kharaulakh section, like in others of the eastern facies zone, with a great delay—near the lower boundary of the Atdabanian Stage. Thus, the U–Pb date of 534 Ma determines the upper, but not the lower boundary of the Tommotian Stage, unless it marks the emplacement time of 40-m-thick dolerite sill overlying conglomerates with dated pebbles (Khomentovsky, 2000). Critical analysis by Khomentovsky also demonstrated an ambiguous significance of the U–Pb age determination obtained for the Meishucun section of China, where the lower boundary of the Tommotian Stage is sufficiently well substantiated from the paleontological standpoint.

Despite the absence of reliable geochronological dates for the base of the Tommotian Stage, the level that separates it from the Nemakit-Daldynian Stage is most appropriate for placing here the lower boundary of the Cambrian System. The important stage in evolution of skeletal fauna that can be seen across this boundary and in chemostratigraphic record implies that the base of the Tommotian Stage is most suitable for worldwide tracing. Unfortunately we have no more space to touch in this article the insurmountable difficulties related with interregional tracing of lower boundaries of the *antiqua* Zone and Nemakit-Daldynian Stage (Khomentovsky and Karlova, 1992; Khomentovsky *et al.*, 1998; Khomentovsky, 2000).

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Reviewer M.A. Fedonkin

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Significance of *Coniopteris* and Czekanowskiales for the Jurassic Stratigraphy of West Siberian Continental Deposits

A. I. Kirichkova*, L. I. Bystritskaya, and T. A. Travina***

**All-Russia Petroleum Research and Geological Prospecting Institute, Liteinyi pr. 39, St. Petersburg, 109104 Russia*

***Tomsk State University, pr. Lenina 36, Tomsk, 634050 Russia*

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Abstract—The detailed taxonomy of Czekanowskiales and genus *Coniopteris* dominating in Jurassic paleoflora of West Siberia, composition and stratigraphic confinement of their assemblages, and species of a restricted stratigraphic range are considered. Combinations of *Coniopteris, Phoenicopsis*, and *Czekanowskia* species characteristic of certain age intervals are first recognized. These combinations may be used for stratigraphic subdivision and correlation of West Siberian continental deposits.

Key words: West Siberia, Jurassic, taphoflora, formation, paleoflora, Czekanowskiales, ferns.

INTRODUCTION

The adopted plan to update geological survey at the scale 1 : 200000 requires a significant refinement of local stratigraphic schemes and a substantiation of spatial correlations. Particularly difficult are age determinations and chronostratigraphic correlation of continental deposits. The Jurassic continental deposits are widespread in West Siberia. They are overlain by a thick cover of younger sediments, being recovered by a system of boreholes. The Jurassic continental sequence of West Siberia is of particular interest because of its oil-and-gas potential, and its detailed lithologic, facies, geochemical, and other characteristics are presented in many works. In accord with decisions of the Interdepartmental Stratigraphic Conference (*Resheniya...*, 1991), the Jurassic continental sequence is subdivided into a number of formations with due account to the structural and facies zoning of West Siberia (Table 1). Ages of lithostratigraphic units distinguished in the Jurassic continental sequence and their correlation are rather controversial however. In this connection, new local stratigraphic schemes, detailed biostratigraphy, and verified paleontological grounds for spatial correlation were published recently (Shurygin, 1986, 1987; Kontorovich, Andrusevich, *et al.*, 1995; Kontorovich, Il'ina, *et al.*, 1995; Shurygin *et al.*, 1995; Il'ina, 1997; Meledina *et al.*, 1997). However, biostratigraphic characteristics of Jurassic subdivisions (Table 1) are based so far on palynological data only (Il'ina, 1985, 1997; Glushko and Sheiko, 1987; Kostesha *et al.*, 1990,1991; Kabanova and Kostesha, 1997). This is a consequence of inadequately studied taxonomy of plant macrofossils frequently occurring in drill cores.

Teslenko (1970) was first to publish list of the Jurassic plant taxa from boreholes drilled in western to southern Siberia and Tuva, mainly in the Kuznetsk, Chulym-Enisei, and Irkutsk basins. The list included 25 taxa from 20 boreholes. Almost all studied plant remains were poorly preserved and likely originated from the Tyumen' Formation of the Middle Jurassic age. *Clathropteris* leaf remains, based on which Teslenko inferred the Early Jurassic age of corresponding sediments, were encountered in the Well Uvat 2-R only (3008–3020 m interval). He noted that gymnosperm leaves are dominant among plant remains and represent 15 of 25 defined taxa. Unfortunately, plant taxa were identified exclusively from the leaf morphology without due regard for their epidermal structure. Accordingly, Teslenko failed to establish true diversity of genera and species in the plant groups predominant in the Siberian paleoflora.

Mogucheva identified a wider spectrum of plants from the Lower–Middle Jurassic sediments recovered by boreholes in West Siberia (Mogucheva, 1990; Kazakov *et al.*, 1997). However, her lists are generalized for the formation ranges, and gymnosperms were again identified without regard for peculiarities of the leaf epidermis structure. Recent investigations (Doludenko and Rasskazova, 1972; Samylina, 1972, 1986; Kirichkova and Samylina, 1989) showed that the epidermal structure is basic indication of taxonomic affinity of gymnosperm leaf remains. Moreover, the works mentioned above have been published without descriptions and illustrations of taxa, and exact position of flora samples in the sections has not been indicated. Accordingly, they cannot be used for the taxonomic analysis of the West Siberian paleoflora and for the stratigraphy of corresponding deposits.

In 1983–1997 Bystritskaya studied plant macrofossils from Jurassic sediments recovered by boreholes in the Tomsk region (Bystritskaya and Tat'yanin, 1983; Bystritskaya, 1992, 1997; Batyaeva and Bystritskaya, 1994). She described and pictured some species of genera *Coniopteris* and *Nilssonia*, and fructifications of genera *Stenorachis* and *Ixostrobus*. She also investigated for the first time the epidermal structure of leaves of several *Phoenicopsis* species. These were first reliable data on flora from buried Jurassic deposits of the West Siberian regions.

By now we collected abundant core samples with plant remains from over 60 boreholes of West Siberia, which often bear phytoleims of ginkgophyte leaves. Remains of *Czekanowskia, Phoenicopsis, Sphenobaiera, Ginkgo*, and *Leptotoma* representatives, which dominate in Jurassic flora of Siberia, have been comprehensively studied using the epidermal cuticular method (Samylina and Kirichkova, 1991; Samylina, 1972; Samylina and Markovich, 1991; Kirichkova and Travina, 1990, 1993a, 1993b; Travina, 1996; Kirichkova, Batyaeva, *et al.*, 1992). The obtained results revealed a considerable species diversity of above genera and a distinct stratigraphic confinement of the recognized taxa, which can be used consequently in stratigraphic correlation and taxonomic analysis. For instance, the time succession of specific taphofloras from Jurassic sediments of West Siberia (Table 1) was distinguished for the first time in the Tomsk area, latitudinal segment of the Ob' River basin, and in reference boreholes of the Urengoi region (Bystritskaya *et al.*, 1992; Kirichkova and Travina, 1995). A similar succession of plant assemblages was established in the section of the superdeep Well Tyumen' TSG-6 (Kirichkova *et al.*, 1999). In addition, Timoshina studied spore and pollen spectra from the same intervals of the well and in the reference sections of the Lower-Middle Jurassic of southern West Siberia, namely, in Well 8500 of the Kansk coal basin (Table 1).

Shown below is the stratigraphic significance of *Coniopteris* and Czekanowskiales for subdivision and correlation of West Siberian continental sequences.

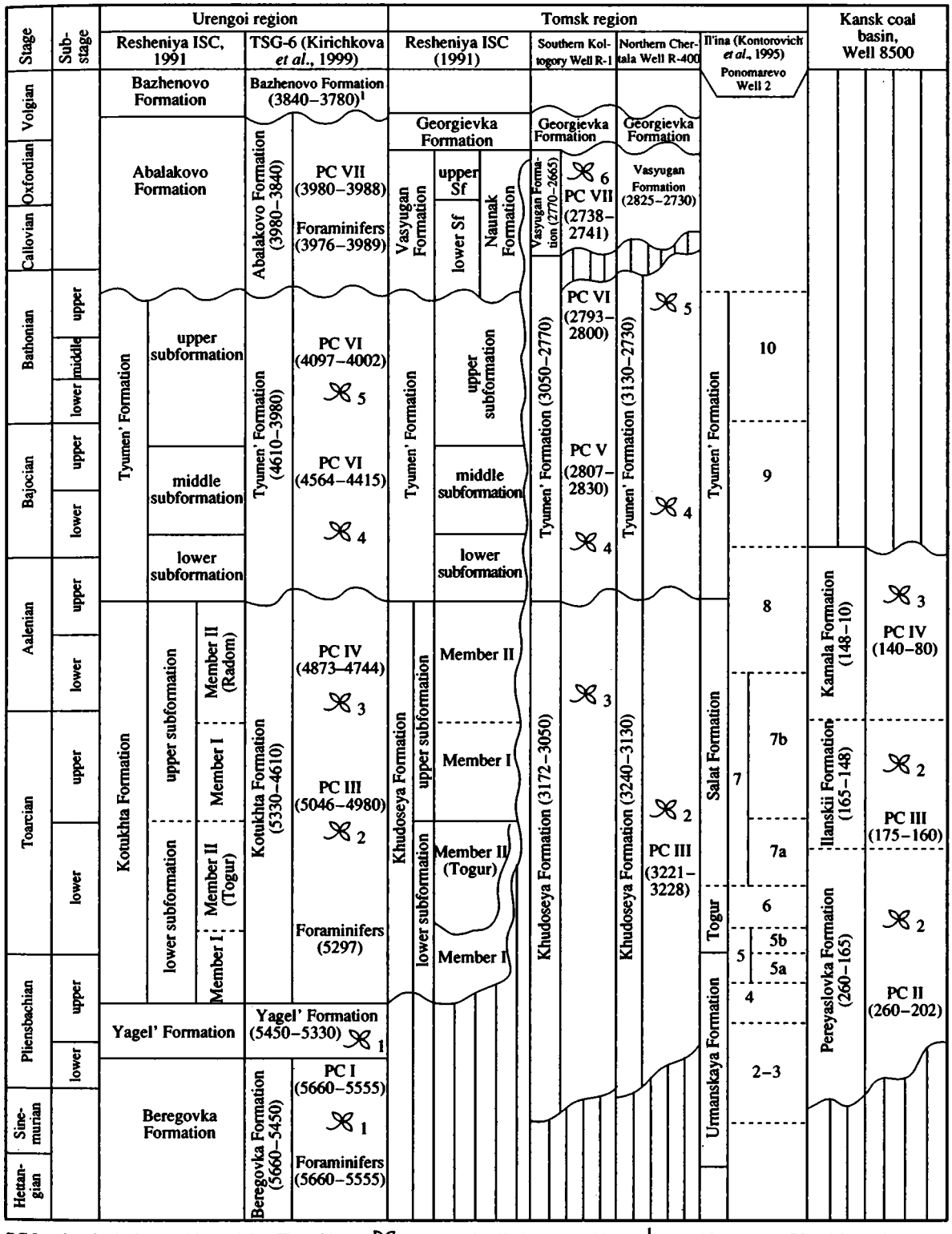
GENUS *CONIOPTERIS*

Pinnate leaves of fern *Coniopteris* are most widespread in the Jurassic continental deposits of West Siberia. Leaf remains of the genus, such as fragments of pinnae, separate pinnules, and scarcer incomplete leaves, occur in cores of many boreholes. A poor preservation degree complicates identification of these remains. In our study, we used therefore a serial sketching of nearly all fragments of sphenopteroid leaves, using a projecting camera. As a result, we managed to determine the variability degree of their morphologic types and to establish the species affiliation of fragmentary remains of pinnate leaves. In general, with regard for records by Teslenko (1970) and Bystritskaya (Bystritskaya, 1992, 1997; Bystritskaya and Tat'yanin, 1983; Batyaeva and Bystritskaya, 1994), the *Coniopteris* species diversity is found to be moderately high in the Jurassic taphofloras of West Siberia. We identified 12 species: *C. burejensis* (Zaless.) Sew., *C. hymenophylloides* (Brongn.) Sew., *C. depensis* E. Lebed., *C. latilobus* Bistrits., *C. levii* Bistrits., *C. kalinoviensis* Bistrits., *C. maakiana* (Heer) Pryn., *C. murrayana* (Brongn.) Brongn., *C. simplex* (Lindl. et Hutt.) Harris, *C.* cf. *spectabilis* Brick, *C. vialovae* Tur.-Ket., and *C. vsevolodii* E. Lebed. Prevailing among *Coniopteris* species are pinnulate forms with strongly dissected laminae. These are *C. simplex, C. vialovae, C. murrayana, C. levii* (Plate I, figs. 12–29), and *C. kalinoviensis. C. maakiana* (Plate I, figs. 30, 31), *C. levii* (Plate I, fig. 17), *C. latilobus* (Plate I, figs. 10, 11), and *C. kalinoviensis* are the West Siberian endemics. The other species are represented by European migrants: *C. hymenophylloides* (Plate I, figs. 1–3), *C. simplex* (Plate I, figs. 12–16), and *C. murrayana* (Plate I, figs. 18–20), or Asian (*C. vialovae* (Plate I, figs. 21–29) and *C.* cf. *spectabilis*) paleofloras. Only *C. burejensis* (Plate I, figs. 4–7), *C. depensis*, and *C. vsevolodii* (Plate I, figs. 8, 9) have connections with the East Siberian paleoflora. As for the stratigraphic confinement of sphenopteroid leaves, their remains are also rather characteristic of the Lower Jurassic beds, e.g., of the lower strata of the Khudoseya Formation and their analogues (Table 1). However, these remains have a poor preservation degree, and it is frequently difficult to identify their species. We identified more or less reliably only *C. hymenophylloides, C. maakiana*, and *C. simplex* (Table 2). In the Lower–Middle Jurassic interval, namely, in the upper part of the Khudoseya Formation and its analogues, the *Coniopteris* leaf remains are considerably more abundant. In addition to the mentioned species, were identified in this part of the section *C. burejensis, C. murrayana*, and *C. vsevolodii* (Table 2). In the lower and middle parts of the Tyumen' Formation, conventionally attributed to the Bajocian, the same species still coexist with abundant *C. levii* and newly appeared *C. depensis* (Table 3). The highest diversity of *Coniopteris* species is established in the second half of the Middle Jurassic. Ten of twelve known species of the genus were distinguished in the upper part of the Tyumen' Formation and its analogues conventionally referred to the Bathonian (Table 4), and *C. simplex, C. burejensis, C. vialovae, C. murrayana*, and *C. depensis* frequent in this interval coexist with *C. latilobus* first appeared here. In the Vasyugan Formation and its analogues, the *Coniopteris* species diversity considerably decreases. *C. latilobus, C. simplex*, and *C. burejensis* are most common species of this unit (Table 5).

CZEKANOWSKIALES

Leaf remains of *Ginkgo* and *Czekanowskia* genera are quite widespread in Jurassic continental sediments

Table 1. Stratigraphic schemes of the Urengoi and Tomsk regions and subdivisions of Jurassic continental deposits in reference borehole sections



PC I, palynological assemblages (after Timoshina); ✂ 1—macrofossil plant assemblages; [1]interval in meters; (Sf) subformation.

Plate I

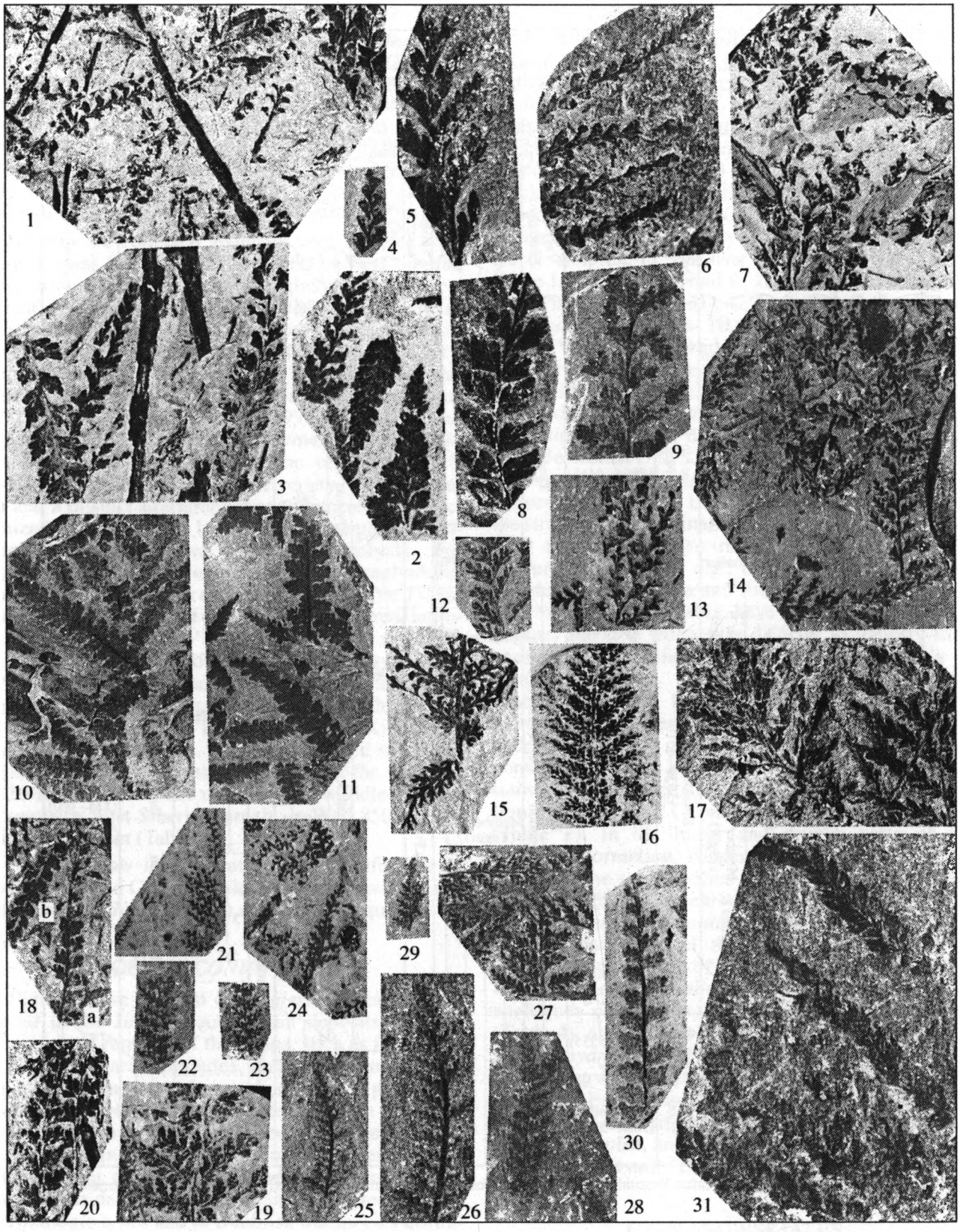

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Table 2. Distribution of the *Coniopteris* species in the Khudoseya Formation and its analogues

Species \ Area, well number, depth, m	Lower subformation						Upper subformation					
	Urengoi 411, 4990–5050	Western Ostaninka 447, 2782–2792	Northern Ostaninka R-1, 2551–2685	Fedyushkino R-2, 3262–3268	Tolparovo R-2, 3123–3130	Var'-Egan p-952, 3245–3254	Urengoi 411, 4714–4726	Urengoi 414, 4250–4267	Urengoi 673, 4001–4016	TSG-6, 4863–4873	Gerasimovka R-8, 2825–2831	Western Luginets R-180, 2475–2499
C. burejensis (Zal.) Sew.									+			
C. hymenophylloides (Brongn.) Sew.		+	+	+		+						+
C. maakiana (Heer) Pryn.					+	+						
C. murrayana (Brongn.) Brongn.									+			
C. simplex (Lindl. et Hutt.) Harris			+			+	+			+		
C. vsevolodii E.Lebed.								+			+	

of West Siberia. Leaves of the latter genus are often strewn over bedding planes of these sediments and represent monotypic burials. Since *Ginkgo* leaves are greatly variable in shape and size, while morphology of Czekanowskiales leaves is not very distinctive in contrast, their taxonomic identification by means of comparative morphological method becomes impossible, and one should use additional diagnostic features, e.g., the leaf epidermal structure. Only the epidermal cuticular method facilitates determination of the generic and specific affiliation of leaf remains, even if they are of a very small size (Plate II, fig. 2a; Plate III, fig. 2a/1).

Plate I. *Coniopteris* species from Jurassic deposits of West Siberia.

(1–3) *Coniopteris hymenophylloides* (Brongn.) Sew., fertile (fig. 1) and sterile pinnae, specimens 858/92 (1) and 858/97 (2), Well Yuzhno-Poselkovaya 1-R, 3058–3063 m; sterile pinnae, specimen 858/99 (3), Well Var'-Egan, p-952, 3233–3245 m; lower and middle parts of the Tyumen' Formation (Bajocian?).

(4–7) *Coniopteris burejensis* (Zaless.) Sew., fragments of fertile (4, 6, 7) and sterile (5) pinnae; specimens 858/72 (4) and 858/68 (5), Bakhilovo Well 126, 2599–2607 m, Vasyugan Formation, uppermost Middle-lowermost Late Jurassic; specimen 858/48 (6), Well Samlat 1-R, 2707–2714 m, uppermost Tyumen' Formation, Middle Jurassic (Bathonian?); specimen 858/59 (7), Well Lesnaya 285, 2630–2645 m, Vasyugan Formation, uppermost Middle–lowermost Upper Jurassic.

(8, 9) *Coniopteris vsevolodii* E.Lebed., fragments of sterile pinnae; specimen 858/261 (8), Well Yuzhno-Tambaevy 77, 2599–2605 m, uppermost Tyumen' Formation, Middle Jurassic (Bathonian?); specimen 858/257 (9), Well Gerasimovka 8, 2825–2831 m, upper part of the Khudoseya Formation, lowermost Middle Jurassic (Aalenian?).

(10, 11) *Coniopteris latilobus* Bistrits., remains of sterile pinnae; specimens 858/111 and 858/112, Verkhne-Zayach'ya Well 80-R, 2688–2695 m, uppermost Tyumen' Formation, Middle Jurassic (Bathonian?).

(12–16) *Coniopteris simplex* (Lindl. et Hutt.) Harris, impressions of fertile (13, 15) and sterile pinnae with heterophyllous strongly dissected pinnules; specimen 858/196 (12), Well Western Venga-Yachin 52, 3208–3220 m, uppermost Tyumen' Formation, Middle Jurassic (Bathonian?); specimen 858/184 (13), Well Var'-Egan p-952, 3245-3254 m, upper part of the Khudoseya Formation, lowermost Middle Jurassic (Aalenian?); specimen 858/205 (14), Yuzhno-Poselkovaya Well 1-R, 2956–2963 m; specimen 858/206 (15), Well Snezhnaya 133, 2491 m, uppermost Tyumen' Formation, Middle Jurassic (Bathonian?); specimen 137/8 (16), Well Archinskaya 40, 2595 m, Vasyugan Formation, uppermost Middle–lowermost Upper Jurassic.

(17) *Coniopteris levii* Bistrits., remains of incomplete sterile pinnate leaves; specimen 116/2, Well Northern Kalinovaya 31, 2835–2842 m, lower and middle parts of the Tyumen' Formation, Middle Jurassic (Bajocian?).

(18a–20) *Coniopteris murrayana* (Brongn.) Brongn., (18b) *Coniopteris latilobus* Bistrits., remains of sterile pinnae, specimen 858/183 (18), Well Southern Talinskaya 20007, 2412–2428 m; specimens 851/181 (19) and 858/176 (20), Well Western Luginets R-180, 2566–2575 m and 2477–2485 m, uppermost Tyumen' Formation, Middle Jurassic (Bathonian?).

(21–29) *Coniopteris vialovae* Tur.-Ket., impressions of small fertile (21, 24) and sterile pinnae; specimen 15 (21, 22, 24), Well Kaimysovy 1-R, 2778–2782 m (collection of Mogucheva, SNIIGGiMS), uppermost Tyumen' Formation, Middle Jurassic (Bathonian?); specimens 858/223 (23) and 858/251 (28), Well Verkhne-Zayach'ya 80-R, 2688–2695 m; specimen 858/250 (25, 26), (26x2) Well Southern Talinskaya 20009, 2471–2482 m; specimen 858/239 (27), Well Lesnaya 285, 2680–2700 m; specimen 858/231 (29), Well Iokhtur 514, 3298–3313 m; uppermost Tyumen' Formation, Middle Jurassic (Bathonian?).

(30, 31) *Coniopteris maakiana* (Heer) Pryn., remains of sterile pinnae; specimen 858/147 (30), Well Tolparovo R-1, 3123–3130 m; specimen 858/151 (31), Well Var'-Egan p-952, 3245–3254 m; lower half of the Khudoseya Formation, Lower Jurassic (conventionally Toarcian).

Table 3. Distribution of the *Coniopteris* species in the lower and middle parts of the Tyumen' Formation

Species \ Area, well number, depth, m	Peredovaya 192, 2671	Southern-Poselkovaya 1-R, 3058–3063	Protochnaya 2, 2471–2478	Var'-Egan 952, 3150-3245	Lysogorskaya 11, 2972–2976	Ostaninka 418, 2617–2524	Kiev-Egan 352, 2760–2772	Urengoi 673, 3950–3960	Northern Kalinovaya 31, 2835–2842
C. burejensis (Zal.) Sew.									
C. depensis E.Lebed.								+	
C. hymenophylloides (Brongn.) Sew.		+		+					
C. levii Bistrits.					+		+		+
C. maakiana (Heer) Pryn.				+					
C. simplex (Lindl. et Hutt.) Harris			+			+			
C. spectabilis Brick									

Species \ Area, well number, depth, m	Northern Urmanskaya 11, 2868–2874	Northern Urmanskaya 11, 2868–2874	Shirotnaya 53, 2976–2981	Northern Chertala 400, 2997–3001	Grushevaya 211, 3141–3161	Lower-Tabaga		Kvartovaya 8, 2716–2723	Gerasimovka 4, 2558
						18, 2729	9, 2766		
C. burejensis (Zal.) Sew.		+				+			
C. depensis E.Lebed.					+				
C. hymenophylloides (Brongn.) Sew.									
C. levii Bistrits.	+		+						
C. maakiana (Heer) Pryn.				+					
C. simplex (Lindl. et Hutt.) Harris							+		
C. spectabilis Brick4								+	+

Using this method, we revealed a considerably higher specific and even generic diversity of Ginkgoales and Czekanowskiales than was suspected earlier. Below we consider species composition and stratigraphic significance of Czekanowskiales represented by genera *Czekanowskia* and *Phoenicopsis*, the major components of the West Siberian paleoflora.

Leaf morphology of *Phoenicopsis* and *Czekanowskia* is very simple. Both genera have simple, 3 to 10-mm-wide (*Phoenicopsis*) or narrower, 0.5 to 4-mm-wide (*Czekanowskia*) dichotomizing leaves, connected at their basis into a cluster and set at the end of a brachyblast (Plate II, fig. 3a; Plate III, fig. 1a). Remains of complete leaves or their clusters are scarce in samples from boreholes; usually found are small fragments of leaves or their segments. Traditionally the complete leaves were identified as *Phoenicopsis angustifolia* Heer or *Ph. speciosa* Heer, depending on the leaf width, while narrow dichotomizing leaves or their segments were attributed to *Czekanowskia rigida* Heer. The more exact identification of such remains follows from their epidermal structure (Plates II–IV).

The Jurassic genus *Czekanowskia* is represented in West Siberia by ten following species: *Cz. auliensis* Dolud., *Cz. baikalica* Kiritch. et Samyl., *Cz. enissejensis* Kiritch. et Samyl., *Cz. irkutensis* Kiritch. et Samyl., *Cz. kanensis* Kiritch. et Samyl., *Cz. obiensis* Kiritch. et Samyl., *Cz. rigida* Heer, *Cz. suntarica* Kiritch. et

Table 4. Distribution of the *Coniopteris* species in the upper part of the Tyumen' Formation and its analogues

Species \ Area, well number, depth, m	Lower-Karbinsk 2, 2824–2827	Urengoi				Urmanskaya 700, 3088–3103	Lysogorskaya 10, 2921–2928	Samlat 1-R, 2707–2714	Lower Tabaga		Southern-Poselkovaya 1-R, 2956–2963	Western Ostaninka 446, 2595–2601	Ostaninka 439, 2593–2599	Western Krasnosel'kup 50, 2961–2976	Western Luginets 180, 2468–2499	Southern Talinskaya			Northern Ostaninka 1, 2578–2585.2	Kalinovaya 17, 2724–2738.7
		411, 3858–3873	336, 3876–3893	423, 3812–3828	282, 3790–3797				18, 2729–2732	12, 2770–2774						20007, 2412–2415	20009, 2471–2482	20010, 2453–2467		
Coniopteris burejensis (Zal.) Sew.	+				+	+	+	+	+		+	+								
C. depensis E.Lebed.		+	+																	
C. hymenophylloides (Brongn.) Sew.														+	+					
C. kalinoviensis Bistrits.																				+
C. latilobus Bistrits.												+			+			+		
C. levii Bistrits.																			+	
C. murrayana (Brongn.) Brongn.										+						+				
C. simplex (Lindl. et Hutt) Harris											+		+							
C. vialovae Tur.-Ket.													+				+			
C. vsevolodii E.Lebed.				+											+					

Species \ Area, well number, depth, m	Middle-Messo-Yakha 2-R, 2487–2499	Northern Chertala 400, 2858–2866	Ety-Pur 175, 3214–3225	Lesnaya 285, 2630–2745	Tolparovo 1-R, 2703–2710	Western Venga-Yakhin 52, 3208–3220	Snezhnaya 133, 2491	Iokhtur 516, 3298–3313	Urmanskaya 4, 2850–2855	Grushevaya 211, 2825–2830	Kaimysovy 1-R, 2778–278	Southern Tambaevy 77, 2599–2605	Verkhne-Zayach'ya 80-R, 2688–2695	Western Ostaninka 447, 2532–2539	Smolyanaya 1-R, 2711–2718	Zapolyarnaya 77, 3851–3862	Northern Urengoi 436, 4130–4138	Kvartovaya 8, 2716–2721	Vesennyaya 249, 2650–2652
Coniopteris burejensis (Zal.) Sew.																			
C. depensis E.Lebed.																+	+		
C. hymenophylloides (Brongn.) Sew.													+					+	
C. kalinoviensis Bistrits.																			
C. latilobus Bistrits.													+	+					+
C. levii Bistrits.																			
C. murrayana (Brongn.) Brongn.	+		+	+	+		+								+				
C. simplex (Lindl. et Hutt) Harris					+	+			+						+				
C. vialovae Tur.-Ket.		+		+				+		+	+		+						
C. vsevolodii E.Lebed.												+							

Table 5. Distribution of the *Coniopteris* species in the Vasyugan Formation and its analogues

Species \ Area, well number, depth, m	Igol'skaya 6, 2780–2787	Bakhilovo 126, 2599–2607	Bakhilovo 127, 2651–2655	Dvoinaya 4-R, 2587–2593	Krylovskaya 1, 2433–2438	Lesnaya 285, 2630–2645	Gerasimovka 8-R, 2582–2588	Koltogory 2-R, 2362–236	Upper-Kambarskaya 294, 2423–2435
C. burejensis (Zal.) Sew.		+		+		+			
C. depensis E.Lebed.								+	
C. hymenophylloides (Brongn.) Sew.						+			
C. latilobus Bistrits.	+		+		+		+		
C. levii Bistrits.									
C. simplex (Lindl. et Hutt.) Harris									+
C. vsevolodii E.Lebed.									

Species \ Area, well number, depth, m	Sel'veikinskaya 2-R, 2545	Urmanskaya 2-R, 2674–2680	Archinskaya 40, 2595	Upper-Kolik-Egan 65, 2601–2608	Tambaevy 2-R, 2490	Pervomaiskaya 272, 2550–2561	Sosnovskaya 2-R, 2503–2412	Vesennyaya 354, 2553–2560	Lower Tabaga 9, 2610–2615
C. burejensis (Zal.) Sew.									+
C. depensis E.Lebed.									
C. hymenophylloides (Brongn.) Sew.								+	
C. latilobus Bistrits.	+					+			
C. levii Bistrits.							+		
C. simplex (Lindl. et Hutt.) Harris	+	+	+	+	+	+		+	+
C. vsevolodii E.Lebed.		+							

Samyl., *Cz. tomskiensis* Kiritch. et Samyl., and *Cz. vera* Kiritch. et Samyl.

The genus *Phoenicopsis* incorporates nine species: *Ph. angustifolia* Heer, *Ph. dentata* Pryn., *Ph. irkutensis* Dolud. et Rasskaz., *Ph. markovitchiae* Kiritch. et M.Shishk., *Ph. samylinae* Kiritch. et Moskv., *Ph. sibirica* Kiritch. et Trav., *Ph. taschkessiensis* Krasser, *Ph. striata* Kiritch. et Trav., and *Ph. varia* Kiritch. et Trav. The analyzed species distribution in the sections revealed their restricted stratigraphic ranges.

Plate II. Characteristic *Czekanowskia* species from the Lower–Middle Jurassic deposits of West Siberia.

(1) *Czekanowskia enissejensis* Kiritch. et Samyl., specimen 82: complete open cuticle of the amphistomatic leaf segment (a), lower (l), side (s), and upper (u) epidermis, ×110; portion of the upper epidermis (b), ×200; stomatal complex (c), ×350; Well Northern Chertala R-400, 3160 m, lower half of the Khudoseya Formation, Lower Jurassic (conventionally Toarcian).

(2) *Czekanowskia irkutensis* Kiritch. et Samyl., specimen 90: fragments of leaf segments (a); complete unrolled cuticle of the amphistomatic leaf segment (b): lower (l), side (s), upper (u) epidermis, ×110; portion of the lower epidermis (c), ×200; Well Fedyushkino R-2, 3268.7–3274.7 m, upper part of the Khudoseya Formation, lowermost Middle Jurassic (Aalenian?).

(3) *Czekanowskia rigida* Heer., specimen 561/64, leaf cluster (a), complete unrolled cuticle of the amphistomatic leaf segment (b): lower (l), side (s), and upper (u) epidermis, ×110; portion of the lower epidermis (c), ×200; stomatal complex (d), ×350; Well Gerasimovka 8-R, 2825–2832 m, lower half of the Khudoseya Formation, uppermost Lower Jurassic (Toarcian?).

Plate II

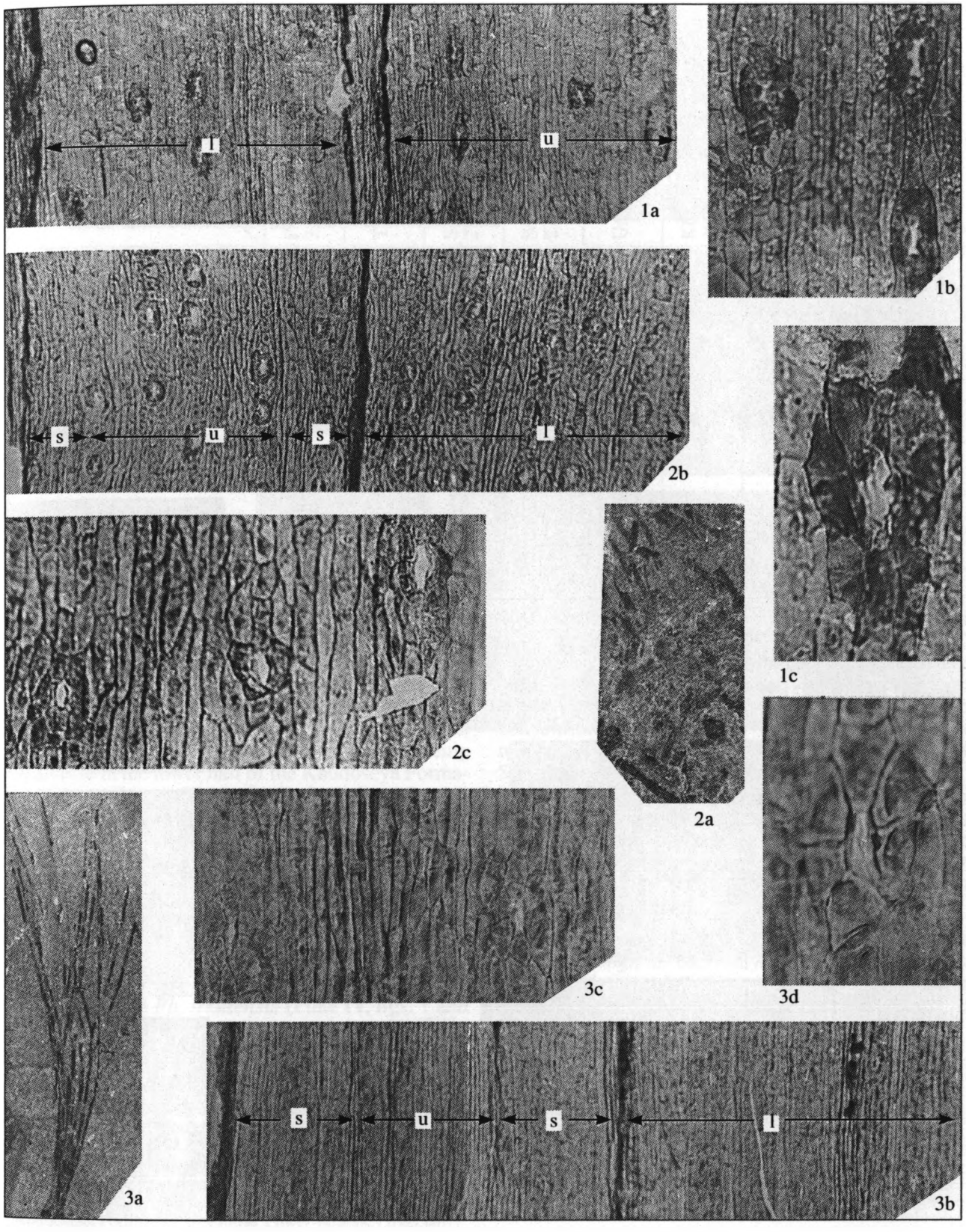

Plate III

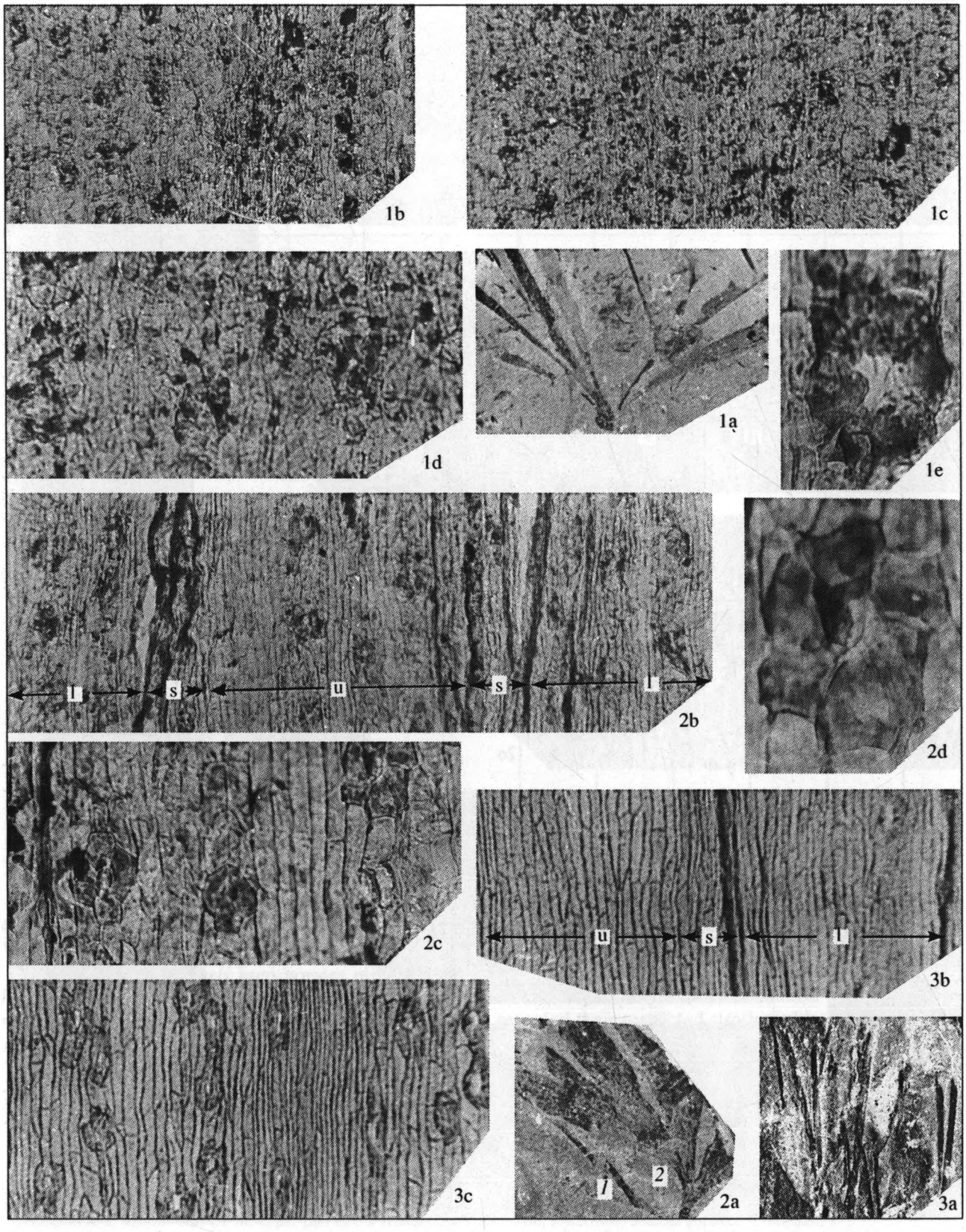

Table 6. Distribution of Czekanowskiales in the lower half of the Khudoseya Formation and its analogues

Species \ Area, well number, depth, m	Prikoltogory R-2, 3160–3168	Talinskaya 129, 2582–2584	Southern Talinskaya R-901, 2760-276	Southern Talinskaya 20012, 2723–2737	Urman 11, 3087–3097	Kharampur 340, 3800–3805	Gerasimovka 8-R, 2825–2832	Northern Chertala R-400, 3113–3160	Southern Chasel'ka R-15, 3742–3753
Phoenicopsis angustifolia Heer			+		+				
Ph. dentata Pryn.								+	
Ph. irkutensis Dolud. et Rasskaz.	+			+					
Ph. striata Kiritch. et Trav.						+			
Cz. baikalica Kiritch. et Samyl.								+	+
Cz. enissejensis Kiritch. et Samyl.								+	
Cz. irkutensis Kiritch. et Samyl.		+							
Cz. obiensis Kiritch. et Samyl.	+								
Czekanowskia rigida Heer		+				+	+	+	

In the Lower Jurassic part of the Khudoseya Formation (lower half) and its analogues, the *Phoenicopsis* leaf remains are scarce, represented by four species (Table 6). Leaves of *Ph. striata* are recorded in this part of the section only; among others, remains of *Ph. irkutensis* and *Ph. angustifolia* are most common. Leaves of *Czekanowskia* represent five species, being most diverse in the lower half of the Khudoseya Formation and its analogues (Table 6). *Cz. baikalica, Cz. obiensis*, and *Cz. rigida* (Plate II, fig. 3; Plate III, figs. 2 and 3), the latter most abundant, were encountered exclusively in the Lower Jurassic beds.

In the upper part of the Khudoseya Formation and its analogues, corresponding to the basal Middle Jurassic, occurrence rates of genera change considerably. More widespread *Phoenicopsis* is represented by six species (Table 7), among which *Ph. angustifolia, Ph. varia*, and *Ph. irkutensis* (Plate IV, figs. 1 and 3) are most common. *Czekanowskia* is represented in this interval by *Cz. irkutensis* (Plate II, fig. 2) and *Cz. enissejensis* (Plate II, fig. 1), the latter is rather abundant (Table 7). Upward in the section, precisely in the lower and middle parts of the Tyumen' Formation conventionally referred to the Bajocian, the total number of *Phoenicopsis* species is the same (Table 8). Simultaneously, *Ph. angustifolia* is missing from this level, *Ph. irkutensis* (Plate III, fig. 1) is scarce, *Ph. samylinae* occurs in the lower Tyumen' Subformation only, and *Ph. sibirica* is newcomer. *Phoenicopsis* leaf remains, mostly those of *Ph. varia, Ph. dentata*, and *Ph. sibirica*, are as common in this interval as below.

In the upper part of the Tyumen' Formation and its analogues, e.g., in the Malyshevka Formation, Czekanowskiales have different abundance and species composition. Genus *Czekanowskia* represented by

Plate III. Characteristic Czekanowskiales species from the Lower-Middle Jurassic deposits of West Siberia.

(1) *Phoenicopsis irkutensis* Dolud. et Rasskaz., specimen 498/56; cluster of linear leaves (a) and upper epidermis of the amphistomatic leaf (b), ×110; lower epidermis (c), ×110; portion of the lower epidermis (d), ×200; stomatal complex (e), ×350; Well Tolparovo R-1, 2696.8–2703.3 m, lower-middle parts of the Tyumen' Formation (Bajocian?).

(2) *Czekanowskia obiensis* Kiritch. et Samyl., specimen 38; leaf segment of Czekanowskiales (a/1), incomplete leaf of *Ginkgo* sp. (a/2); complete unrolled cuticle of the amphistomatic leaf segment: lower (l), side (s), and upper (u) epidermis, ×110; portion of the lower epidermis (c), ×200; stomatal complex (d), ×350; Well Prikoltogory R-2, 3160–3168 m, lower half (Togur sequence) of the Khudoseya Formation, uppermost Lower Jurassic (Toarcian?).

(3) *Czekanowskia baikalica* Kiritch. et Samyl., specimen 7; fragments of leaf segments (a) and portion of the upper and side epidermis of the amphistomatic leaf (b), ×110; portion of the lower epidermis with stomatal stripes (c), ×200; Well Southern Chasel'ka R-15, 3742–3753 m, lower half of the Kotukhta Formation, uppermost Lower Jurassic (Toarcian?).

Plate VI

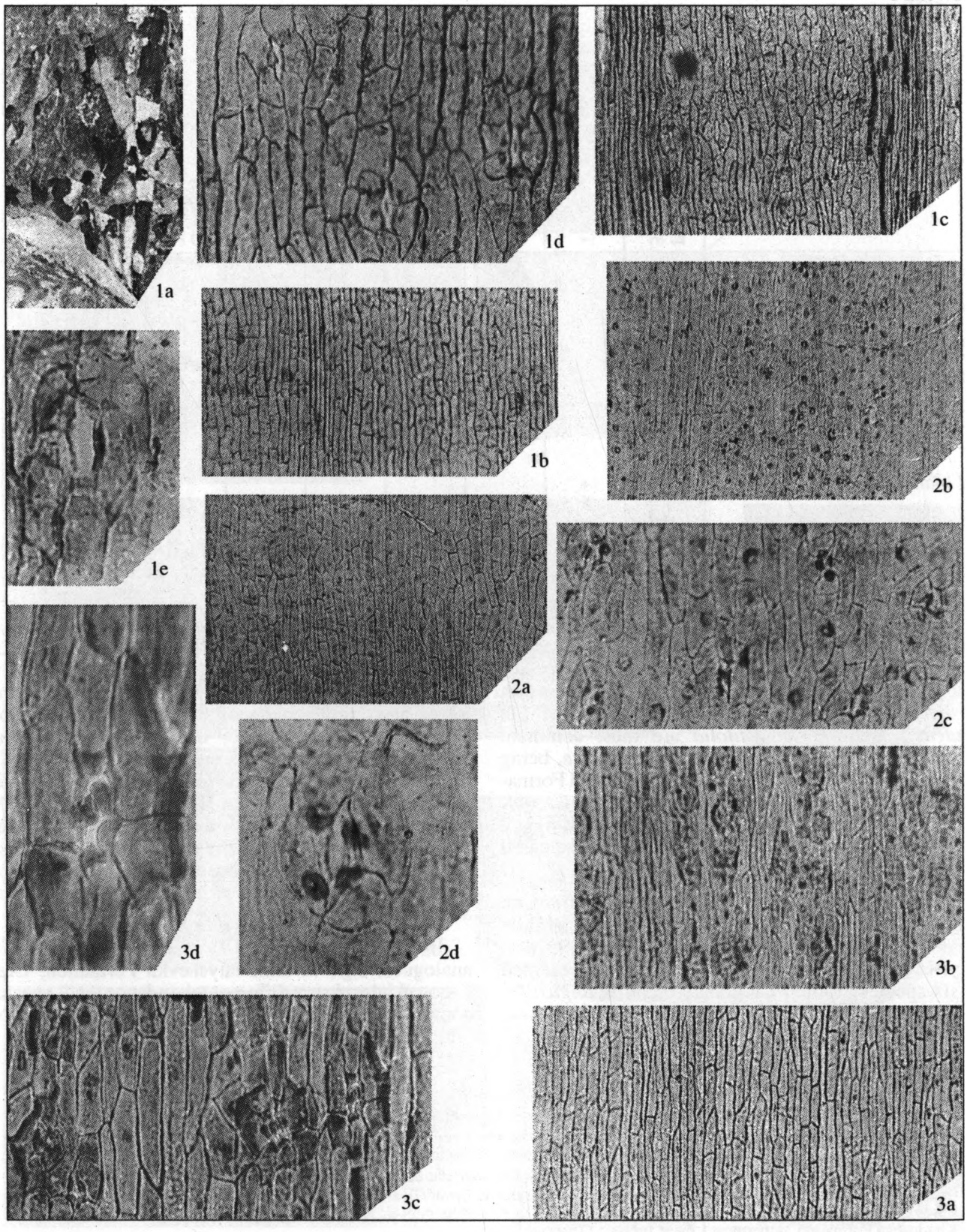

Table 7. Distribution of Czekanowskiales in the upper part of the Khudoseya Formation and its analogues

Species \ Area, well number, depth, m	Fedyushkino R-2, 3262.7–3274.7	Western Karai 2-R, 3166–3172	Potanaiskaya 75-R, 2244–2256	Lontyn-Yakh 59, 2655	Yakhlinskaya 49, 2604–2620	Gerasimovka					Northern Ostaninka	
						11, 2800–2810	10, 2809–2816	19, 2777–2781	7, 2699–2705	17, 2762–2768	5-R, 2744–2750	7-R, 2763–2772
Phoenicopsis angustifolia Heer	+	+	+	+	+							
Ph. dentata Pryn.												
Ph. irkutensis Dolud. et Rasskaz.	+											
Ph. markovitchiae Kiritch. et M. Shishk.												
Ph. samylinae Kiritch. et Moskv.								+	+			
Ph. varia Kiritch. et Trav.						+	+			+	+	+
Cz. enissejensis Kiritch. et Samyl.												
Czekanowskia irkutensis Kiritch. et Samyl.	+											

Species \ Area, well number, depth, m	Smolyanaya 1-R, 2917–2924	Tolparovo 1-R, 2790–3180	Lyantorskaya 301, 3107–3121	Talinskaya		Var'-Egan 952, 3220–3227	Western Novogodnyaya 210, 3597–3610	Western Ostaninka	
				6079, 2050–2057	129, 2501–2513			447, 2785	446, 2520–2523
Phoenicopsis angustifolia Heer									
Ph. dentata Pryn.			+						
Ph. irkutensis Dolud. et Rasskaz.	+	+							
Ph. markovitchiae Kiritch. et M. Shishk.								+	
Ph. samylinae Kiritch. et Moskv.								+	
Ph. varia Kiritch. et Trav.	+		+						
Cz. enissejensis Kiritch. et Samyl.		+			+	+	+		+
Czekanowskia irkutensis Kiritch. et Samyl.		+		+					

Plate IV. Characteristic *Phoenicopsis* species from the Jurassic deposits of West Siberia.

(1) *Phoenicopsis varia* Kiritch. et Trav., specimen 15; leaf fragments (a) and upper epidermis of a hypostomatic leaf (b), ×110; lower epidermis with stomatal stripes (c), ×110; portion of the lower epidermis (d), ×200; stomatal complex (e), ×350; Well Gerasimovka 17, 2762–2768, upper part of the Khudoseya Formation, initial Middle Jurassic (Aalenian?).

(2) *Phoenicopsis sibirica* Kiritch. et Trav., specimen 43; upper epidermis of a hypostomatic leaf (a), ×110; lower epidermis with a stomatal stripe and large papillae in common epidermal cells (b), ×110; portion of the lower epidermis (c), ×200; stomatal complex (d), ×350; Well Southern Tambaevy 76, 2650.7–2657.7 m, Vasyugan Formation, uppermost Middle-lowermost Upper Jurassic.

(3) *Phoenicopsis angustifolia* Heer., specimen 4; upper epidermis of a hypostomatic leaf (a), ×110; lower epidermis with stomatal stripes (b), ×110; portion of the lower epidermis (c), ×200; stomatal complex (d), ×350; Well Potanaiskaya 75-R, 2244–2256 m, upper part of the Kotukhta Formation, lowermost Middle Jurassic (Aalenian?).

Table 8. Distribution of Czekanowskiales in the lower and middle parts of the Tyumen' Formation

Species \ Area, well number, depth, m	Talinskaya: 20007, 2520–2534	Talinskaya: 2797, 2537–2547	Puglalymskaya 90, 2690–2697	Var'-Egan 952, 3110–3150	Krasnoleninskaya R-805, 2657–2673	Alekhino R-3051, 3095–3110	Kuinlorskaya 262, 2830–2842	Upper-Zayach'ya R-80, 2719–2872
Phoenicopsis dentata Pryn.	+	+						+
Ph. irkutensis Dolud. et Rasskaz.								
Ph. markovitchiae Kiritch. et M. Shishk.								
Ph. samylinae Kiritch. et Moskv.	+		+					
Ph. sibirica Kiritch. et Trav.								
Ph. varia Kiritch. et Trav.				+	+	+	+	
Czekanowskia enissejensis Kiritch. et Samyl.		+						

Species \ Area, well number, depth, m	Northern Kalinovaya: 26, 2885–2890	Northern Kalinovaya: 31, 2814–222	Gerasimovka 3, 2547–2753	Western Luginets 180, 2468–2477	Srednesogamskaya 50, 2669–2683	Yakhlinskaya 49, 2347–2360	Kalinovaya 9, 2788–2795	Southern Koltogory 1-R, 2976–3047	Tolparovo R-1, 2696.8–2703.3
Phoenicopsis dentata Pryn.	+								
Ph. irkutensis Dolud. et Rasskaz.									+
Ph. markovitchiae Kiritch. et M. Shishk.			+						
Ph. samylinae Kiritch. et Moskv.								+	
Ph. sibirica Kiritch. et Trav.		+				+		+	
Ph. varia Kiritch. et Trav.									
Czekanowskia enissejensis Kiritch. et Samyl.				+	+		+		

other species (Table 9), holds the leading position in species diversity and occurrence frequency of leaf remains in sediments. the newly appeared species are *Cz. kanensis, Cz. tomskiensis*, and *Cz. vera*. Among *Phoenicopsis* forms, *Ph. sibirica* and *Ph. varia* became widespread.

Czekanowskiales from the Vasyugan Formation and its analogues are almost of the same composition (Table 10). Genus *Phoenicopsis* is represented by abundant *Ph. sibirica* (Plate IV, fig. 2) associated with scarce *Ph. dentata* and *Ph. markovitchiae. Czekanowskia* forms are more diverse, represented by abundant *Cz. tomskiensis* and scarcer *Cz. vera* and *Cz. auliensis.*

Thus, new taxonomic data on Czekanowskiales, one of the dominant groups of West Siberian paleoflora, and on *Coniopteris* ferns are encouraging. They demonstrated a considerably greater species diversity of Czekanowskiales, outlined the range of morphological variability of *Coniopteris* sphenopteroid leaves, and refined the species composition of the genus. The subsequent analysis of Czekanowskiales and *Coniopteris* species distribution in the section revealed their more restricted stratigraphic range. The obtained results open a possibility to use with confidence the studied plant groups for stratigraphic subdivision of continental sequences and emphasize stratigraphic importance of certain taxa.

Table 9. Distribution of Czekanowskiales in the upper part of the Tyumen' Formation and its analogues

Species \ Area, well number, depth, m	Tolparovo 1-R, 2690–2730	Urmanskaya 4-R, 2678–2680	Igol'skaya 24-R, 2760–2800	Lower-Tabaga 21, 2647–2654	Kalinovaya 31, 2814–2822	Northern Kalinovaya 29, 2619–2624	Northern Kalinovaya 9, 2788–2795	Lyantorskaya 304, 2863-2877	Khancheiskaya 663, 2960–2975	Malyi Agan 904, 2853–2866	Malyi Agan 901, 2888–2895	Middle-Mes-so-Yakha 2-P, 2499–2511	Middle-Mes-so-Yakha 4-P, 2870–2880
Ph. dentata Pryn.													
Phoenicopsis sibirica Kiritch. et Trav.	+	+		+		+		+	+				
Ph. varia Kiritch. et Trav.										+	+	+	+
Cz. auliensis Dolud.													
Cz. cf. *enissejensis* Kiritch. et Samyl.													
Cz. kanensis Kiritch. et Samyl.												+	
Cz. tomskiensis Kiritch. et Samyl.													
Czekanowskia vera Kiritch. et Samyl.													

Species \ Area, well number, depth, m	Western Krasnosel'kup 46, 2561–2570	Stolbovaya 77, 2622–2626	Northern Dem'yanka 20, 2994–2997	Talinskaya 5561, 2515–2578	Western Venga-Yakhinskaya 52, 3440–3455	Western Luginets 180, 2477–2566	Lesnaya 285, 2680–2700	Northern Kashatskaya 33, 2559–2572	Koshil'skaya 304, 2650–2657	Kholmistaya 663, 2960–2975	Ostaninka 451, 2468–2471
			+	+							
Phoenicopsis sibirica Kiritch. et Trav.										+	+
Ph. varia Kiritch. et Trav.	cf.	+									
Cz. auliensis Dolud.			+						+		
Cz. cf. *enissejensis* Kiritch. et Samyl.							+	+			
Cz. kanensis Kiritch. et Samyl.			+								
Cz. tomskiensis Kiritch. et Samyl.					+	+					
Czekanowskia vera Kiritch. et Samyl.	+										

Table 11 shows that the West Siberian sediments of the second half of the Early Jurassic are characterized by a low diversity of genus *Coniopteris*, main species of which are the widespread *C. hymenophylloides* and scarcer *C. simplex*. Czekanowskiales of that time were more diverse. Their common forms were *Czekanowskia rigida, Phoenicopsis angustifolia* and *Ph. irkutensis*. The beginning of the Middle Jurassic (conventionally the Aalenian) is marked by changes in both plant groups, though diversity of *Coniopteris* ferns was still insignificant. In addition to formerly known species, *C. vsevolodii, C. maakiana, C. murrayana,* and *C. burejensis* appeared in this group. Genus *Phoenicopsis* had the greatest species diversity at that time and dominated among Czekanowskiales in the taphofloras. It was represented by *Ph. angustifolia, Ph. dentata, Ph. varia* and *Ph. samylinae* (Table 11). Genus *Czekanowskia* was less diversified; its spe-

Table 10. Distribution of Czekanowskiales in the Vasyugan Formation and its analogues

Species \ Area, well number, depth, m	Bakhilovo 12, 2645–2647	Southern Koltogory 1-R, 2740–2748	Gerasimovka 8, 2425–2466	Gerasimovka 3, 2547–2553	Gerasimovka 1, 2584–2588	Upper-Zayach'ya 80-R, 2688–2695	Southern Chasel'ka 15-R, 2840–2866	Western Ostaninka 447, 2390–2400	Ugut 61, 2876-2883	Pervomaiskaya 272, 2550–2561	Koshil'skaya 304-R, 2330–2337	Kalinovaya 10, 2567–2571	Southern Ugut 46, 2959–2963	Western Luginets 180, 2424–2427	Southern Tambaevy 76, 2650.7–2657.7
Ph. dentata Pryn.											+				
Ph. markovitchiae Kiritch. et M.Shishk.				+											
Phoenicopsis sibirica Kiritch. et Trav.					+	+	+	+		+	+				+
Cz. auliensis Dolud.	+								+						
Cz. cf. *enissejensis* Kiritch. et Samyl.				+											
Cz. suntarica Kiritch. et Samyl.							+								
Czekanowskia tomskiensis Kiritch. et Samyl.				+				+				+	+	+	
Cz. vera Kiritch. et Samyl.	+	+													

cies *Cz. rigida* and *Cz. baikalica* disappeared, *Cz. irkutensis* still existed, and *Cz. enissejensis* gained a higher abundance. In the mid-Middle Jurassic time (conventionally in the Bajocian), the diversity of Czekanowskiales and *Coniopteris* forms significantly decreased in the West Siberian taphofloras. At that time, *Phoenicopsis varia* and *Ph. dentata* were characteristic forms among Czekanowskiales; *Coniopteris levii* and *C. maakiana* dominated among *Coniopteris* species. *C. burejensis* and *C. simplex* became more common than before. Substantial changes in composition of these plant groups are recorded in taphofloras from the upper part of the Tyumen' Formation and its analogues (conventionally of the Bathonian Stage). In this period, genus *Coniopteris* was most diverse in composition. The most common species from this part of the section are *C. burejensis, C. murrayana, C. depensis, C. vialovae, C. simplex*, and newly appeared *C. latilobus*. Genus *Czekanowskia* experienced diversification and was represented by new species *Cz. tomskiensis, Cz. kanensis*, and *Cz. suntarica*. Only *Cz. enissejensis* survived among all the earlier species (Table 11). *Phoenicopsis* forms common in taphofloras are *Ph. varia* and *Ph. sibirica*. The species diversity of both *Coniopteris* and Czekanowskiales groups declined sharply in the terminal Middle–initial Late Jurassic time (conventionally in the Callovian–Oxfordian). *C. latilobus, C. simplex*, and *C. burejensis* together with *Phoenicopsis sibirica, Czekanowskia tomskiensis, Cz. vera* and *Cz. auliensis* represent the survived species.

Thus, the detailed investigation of fossil plants, which dominated in the Jurassic paleoflora of West Siberia, showed a considerably greater taxonomic diversity of ginkgophytes and of Czekanowskiales among them. Our study also refined diagnostics of sphenopteroid fern species of the genus *Coniopteris*. We managed as well to distinguish species of restricted stratigraphic ranges, which can be used for the detailed subdivision of the Jurassic continental sequences in West Siberia, their spatial correlations, and age determination of corresponding stratigraphic levels.

Reviewers M.P. Doludenko and M.A. Akhmet'ev

Table 11. Distribution of *Coniopteris* (C.), *Czekanowskia* (Cz.), and *Phoenicopsis* (Ph.) species in Jurassic deposits of West Siberia

Species	J_1^2	J_2^1	J_2^2	J_2^3	J_2^4–J_3^1
C. vsevolodii E.Lebed.		2	- - - -		1
C. hymenophylloides (Brongn.) Sew.	4	1	2	5	2
C. levii Bistrits.			5	1	1
C. simplex (Lindl. et Hutt.) Harris	2	2	3	15	18
Ph. striata Kiritch. et Trav.	1				
Ph. dentata Pryn.	1	2	4	2	1
Ph. irkutensis Dolud. et Rasskaz.	2	3	1		
Ph. angustifolia Heer	2	5			
Cz. baikalica Kiritch. et Samyl.	1				
Cz. irkutensis Kiritch. et Samyl.	1	3			
Cz. obiensis Kiritch. et Samyl.	1				
Cz. enissejensis Kiritch. et Samyl.	1	5	5	2	1
Cz. rigida Heer	5				
C. burejensis (Zal.) Sew.		1	3	8	10
C. maakiana (Heer) Pryn.	2	2	2		
C. murrayana (Brongn.) Brongn.		1	- -?- -	8	
Ph. samylinae Kiritch. et Moskv.		3	3		
Ph. varia Kiritch. et Trav.		7	4	6	
Ph. markovitchiae Kiritch. et M. Shishk.		1	1	- -?- -	1
Ph. sibirica Kiritch. et Trav.			2	6	7
C. depensis E.Lebed.			2	4	1
Ph. taschkessiensis Krasser				1	
C. latilobus Bistrits.				5	10
C. spectabilis Brick			2		
C. vialovae Tur.-Ket.				8	
C. kalinoviensis Bistrits.				1	
Cz. vera Kiritch. et Samyl.				1	2
Cz. tomskiensis Kiritch. et Samyl.				2	5
Cz. auliensis Dolud.				2	2
Cz. kanensis Kiritch. et Samyl.				3	
Cz. suntarica Kiritch. et Samyl.					1

(1) Amount of boreholes, which yield a taxon, data by Bystritskaya (1983, 1992) inclusive.

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Jurassic and Cretaceous Sea Basins in the Southeastern Part of the Lesser Caucasus: Sedimentation Settings and Foraminiferal Assemblages

A. I. Voznesenskii*, T. N. Gorbachik, and K. I. Kuznetsova***

**Geological Institute, Russian Academy of Sciences, Pyzhevskii per. 7, Moscow, 109017 Russia*

*** Moscow State University, Vorob'evy gory, Moscow, 119899 Russia*

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Abstract—Structural features and composition of Jurassic deposits in the Kafan tectonic block that is located in the southeastern part of the Lesser Caucasus are considered. Three carbonate members bearing foraminifers are distinguished in the Upper Jurassic–Valanginian sequence of volcanogenic–sedimentary rocks of the Zangezur Group. The overlying Hauterivian–middle Aptian stratigraphic interval is also subdivided into three carbonate members that yield foraminifers. Changes in composition of foraminiferal communities and in lithological features are established below the base of the Valanginian Stage and at the appearance level of orbitolinids within the Barremian interval. As is shown, the lower (Oxfordian–Tithonian) and upper (Valanginian–Aptian) parts of the whole sequence accumulated in different paleolandscape settings. The analyzed paleocoenoses suggest that there were two types of foraminiferal communities: the older (Oxfordian–Tithonian) one with dominant cosmopolitan genera and the subsequent population with prevailing endemic genera of Tethyan affinity. The facies structure of investigated sequence and compositions of foraminiferal assemblages are used to elucidate the evolution history of sea basin that existed in the southeastern part of the Lesser Caucasus during the Late Jurassic–Early Cretaceous time.

Key words: foraminifers, endemics, cosmopolitan genera, Jurassic to Cretaceous deposits, facies zones, paleolandscape settings, carbonate and terrigenous rocks.

Accumulation environments of Jurassic and Cretaceous deposits in the Lesser Caucasus and peculiar features of biota that inhabited sea basins of that time were under keen attention of researchers during several last decades. Paradoxical as it may seem, but the Jurassic–Cretaceous regional sequence is inadequately studied so far and insufficiently characterized in publications. This remark concerns not only the fossil microfauna that was used for stratigraphic subdivision of deposits in question, but also the sequence itself that is of variable completeness and heterogeneous in facies aspects even within a single Kafan block. Among works directly devoted to investigation of that block, we should mention publications by Akopyan (1962), Azaryan (1982), Gorbachik *et al.* (1987), Leven (1985), Leven and Uspenskaya (1990). In their work, Gorbachik and her colleagues used foraminifers to outline the regional Jurassic to Cretaceous stratigraphy and to facilitate its compatibility with other available results. As they established, there is a number of unsolved questions, and we turn back to this problem here, reconsidering it in a more detailed manner. As a rule, any reconsideration is expedient when one obtains new results or newly collected paleontological materials, etc. In the current politic situation, new field observations in the mentioned region are impossible, and we followed another way. First, we revised all materials collected in abundance during the field seasons of the 1980s in the Kafan region. Second, we extended the stratigraphic range under examination up to the upper Aptian Substage of the Lower Cretaceous. Third, we resorted to the help of lithological-facies and sedimentological analyses in order to elucidate the evolution history of paleobasins and paleolandscape environments of sedimentation.

The complicated sequence of Jurassic to Lower Cretaceous deposits was studied in a dozen of individual sections, completeness of which is variable. Foraminifers and lithological peculiarities of rocks were examined in several hundreds of thin sections. Foraminifers are irregularly distributed throughout the sequence: they are most abundant in the Tithonian, Barremian, and Aptian intervals, whereas the Berriasian part composed of volcanic rocks is barren of these fossils. Limestones of variable composition, which represent in the sequence "marine precipitates proper," are resistant against the decomposition techniques, and we identified the foraminiferal taxa based on the test sections visible in petrographic preparations.

Carbonate rocks studied under microscope are described in terms of the Danham's classification (Wilson, 1980) that characterizes in detail composition,

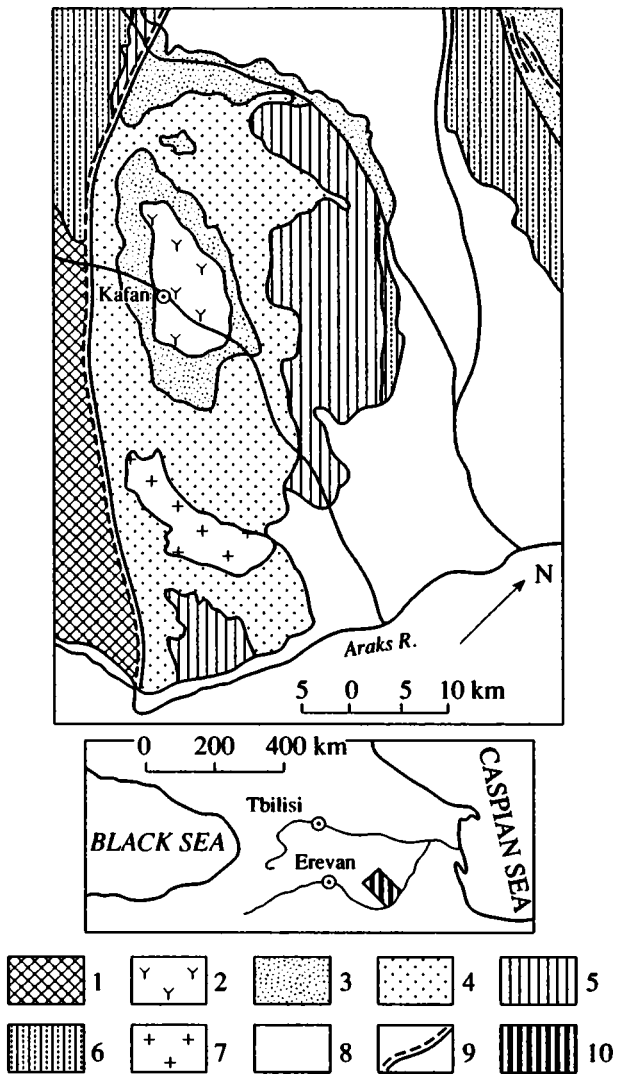


Fig. 1. Schematic geological map of the Kafan region: (1) Paleozoic metamorphic rocks; (2) Bajocian volcanic rocks; (3) lower Oxfordian–Kimmeridgian and (4) Tithonian–Valanginian volcanic, volcanogenic–sedimentary, and carbonate rocks; (5) Hauterivian–Albian and (6) Upper Cretaceous volcanogenic–sedimentary and carbonate deposits; (7) granite and granodiorite intrusions; (8) Cenozoic sediments; (9) large regional faults; (10) study area.

quantitative proportions, and size ranges of their clastic fractions and the type of cementing mass.

GEOLOGICAL STRUCTURE OF THE KAFAN ANTICLINORIUM

The Kafan block in the southeast of the Lesser Caucasus is bounded by two ophiolitic sutures. These are the Sevan–Akera belt on the northeast and the Zangezur belt on the southwest, which separate the block from the Karabakh (Sokolov, 1977) and Erevan–Ordubad (Satian, 1984) tectonic zones, respectively (Fig. 1).

The block corresponds to a large, insignificantly faulted anticlinal structure that has been termed once as the Kafan anticlinorium (*Geologiya SSSR...*, 1970). The anticlinorium core (volcanic complex) is composed of thick, basic to intermediate volcanics of the Aalenian–Bajocian age. The eroded surface of volcanic rocks is overlain by the Upper Jurassic–Lower Cretaceous sequence of volcanic, volcanogenic–sedimentary, and sedimentary rocks, which are about 3 km thick in total. In its structure, composition, and age, this sequence is very close or even identical to the so-called "volcanogenic–carbonate formation" of the Somkhet–Karabakh zone distinguished in the southern margin of the Transcaucasian massif (Mandalyan, 1979).

In the Kafan region, the Upper Jurassic–Valanginian part of the volcanogenic–sedimentary complex was distinguished as the Zangezur Group (Azaryan, 1982). The group is composed in general of basaltic and andesite–basaltic lava flows intercalated with variable pyroclastic and volcanogenic–sedimentary rocks (tuffs, tuffites, tuff-breccias, tuffoconglomerates, and tuffaceous sandstones). Some of intercalations correspond to interlayers of terrigenous and carbonate deposits, which are locally as thick as few dozens of meters and always contain admixture of tuffaceous or volcaniclastic material. In such a case, limestone lentils or interlayers of variable thickness alternate with beds of terrigenous and volcanogenic–sedimentary origin (Fig. 2). Since limestone intercalations are concentrated at several stratigraphic levels, we distinguished a number of corresponding carbonate members within the Upper Jurassic–Lower Cretaceous sequence of the Kafan block. The first member is basal one in the Upper Jurassic sequence and rests on the eroded surface of the Aalenian–Bajocian volcanic rocks. It is composed of reddish brown volcaniclastic large-pebbled conglomerates and sandstones, which enclose lentils of tuffaceous and echinoderm limestones. These rocks fill in depressions of the erosion relief and are up to 150 m thick. The second member at the base of the Tithonian interval is of a complex structure, showing the lenticular alternation of tuffaceous sandstones, tuffites, basaltic lava flows, and yellowish gray sandy–clayey limestones of organogenic-detrital type. It is as thick as few to 250 m. The Valanginian interval of the sequence corresponds almost entirely to the third member that is composed of coarse-grained tuffaceous or volcaniclastic, poorly sorted sandstones intercalated with gray limestones variably enriched in clayey or sandy admixture. The basal sandstone horizon of this member is up to 60 m thick and encloses intercalations of conglomerate-breccias. The total thickness of the member is up to 120 m. In addition to three indicated members, there are individual lentils of sandy organogenic-detrital limestones, which occur at different levels of the Oxfordian–Kimmeridgian interval of the Kafan sequence and are up to 5 m thick and dozens to few hundreds of meters long.

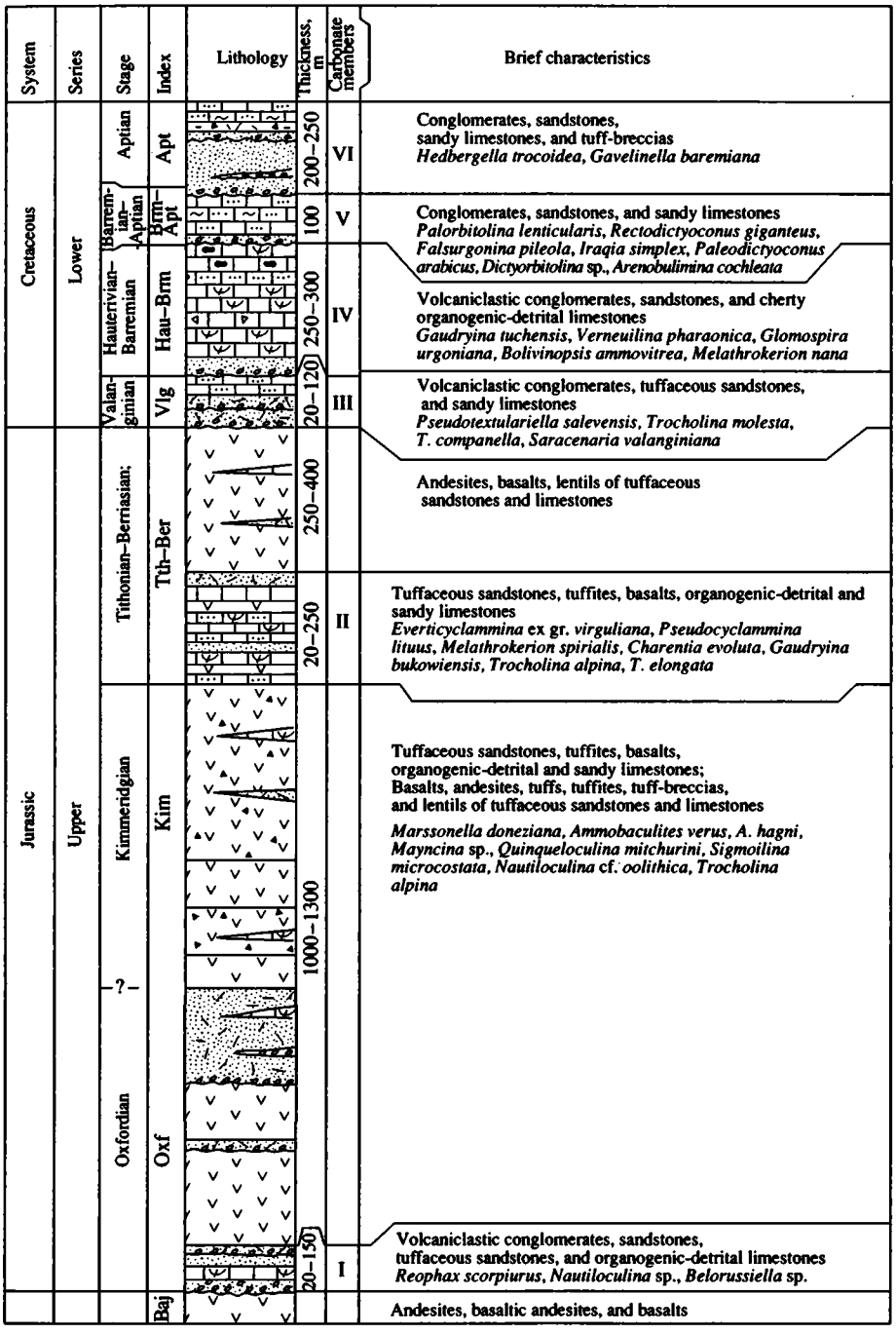


Fig. 2. Litho-stratigraphic column of Upper Jurassic–Lower Cretaceous deposits in the Kafan anticlinorium: (1) andesite, basaltic andesite, or basalt; (2) andesitic and basaltic breccias; (3) tuff-breccia; (4) tuffaceous sandstone; (5) conglomerate; (6) sandstone or gravelstone; (7) organogenic-detrital, (8) sandy, (9) cherty, and (10) brecciated limestones. Stratigraphic indices are given according to Gradstein and Ogg (1996).

The Upper Jurassic–Valanginian sequence is overlain, with insignificant signs of erosion, by the Hauterivian–Barremian terrigenous and carbonate rocks, which represent the fourth member up to 300 m thick. At the base of this member, there is a distinct bed (30–40 m thick) of calcareous volcaniclastic sandstones, gravelstones, and conglomerates, which exhibit rhythmical alternation with sandy limestone lentils marking tops of the rhythms. The middle and upper parts of the Hauterivian–Barremian member are composed of gray

to pinkish gray sandy or clayey limestones of organogenic-detrital type, which bear sometimes the cherty nodules.

The Barremian–Aptian carbonate–terrigenous deposits up to 100 m thick rest with erosion marks, though without angular unconformity, on the top of fourth member. In the lower portion, they are represented by brown basal conglomerate beds (up to 1–3 m thick), which are composed of size-variable pebbles of volcanic rocks and enclose interbeds of calcareous tuffaceous to volcaniclastic sandstones, sandy to clayey limestones, and marls. This 100-m-thick interval corresponds to the fifth carbonate member.

The distinct top boundary of Aptian sandy to calcareous deposits is overlain with erosion marks by volcaniclastic conglomerates, gravelstones, and sandstones, which are up to 250 m thick in total and variably enriched in calcareous and tuffaceous material. The sixth carbonate member is distinguished in the upper portion of the Aptian interval that encloses thin lentils or more extended interbeds of sandy limestones, which are sometimes highly enriched in organogenic detritus.

In the north of the Kafan block, the Upper Jurassic–Lower Cretaceous sequence of volcanogenic–sedimentary rocks is overlain with erosion marks by the Upper Cretaceous sequence of a complex structure that is composed of terrigenous, volcanogenic, and carbonate deposits. In terms of their genesis, all volcanogenic–sedimentary and sedimentary deposits of the Upper Jurassic–Lower Cretaceous sequence accumulated in the mobile coastal, deltaic, and open marine settings of a sea basin, where the dissected volcanic relief characterized the near-coast and deeper shelf zones, the latter more remote from the coast and having less active hydrodynamic parameters. Some volcanic islands could exist as well within the basin. In addition to deposits, which accumulated owing to volcanic activity of terrestrial and submarine volcanoes (lava flows, tuffs, sediments of temporary streams, etc.), the rock complex under consideration includes also the marine deposits proper. First of all, these are sediments of coastal and extremely shallow sea zones, where such landscape elements as beaches, sand tongues, bars, shallow deltas, interdelta areas, carbonate banks, and shallow coastal to isolated reefs can be distinguished. Moreover, sediments of some zones more advanced seaward, e.g., of small depressions with calm hydrodynamics, sporadically occur in the sequence as well.

Deep-water clayey, siliceous, or carbonate–siliceous sediments, which were distinguished by Vishnevskaya (1984) among coeval deposits in the Sevan–Akera zone, are missing from the Upper Jurassic–Lower Cretaceous sequence of the Kafan region, where rocks are extremely depleted in the clayey material.

CHARACTERIZATION OF CARBONATE MEMBERS AND THEIR ACCUMULATION SETTINGS

The first carbonate member at the base of the Upper Jurassic sequence of the Kafan region is of a three-member structure. Its own base is composed of gray-brown massive large-pebbled conglomerate beds, which enclose isolated boulders up to 0.2–0.6 m in diameter. Pebbles of volcanic rocks and volcaniclastic sandstones are moderately sorted, well rounded, having brown ferruginous coating. They are cemented by brown matrix of clayey sandstone. Conglomerate beds are intercalated with brownish green interbeds (up to 0.5 m thick) of massive coarse- to medium-grained volcaniclastic sandstone, which sporadically show presence of isolated small pebbles. These basal conglomerates and sandstones are from 15 to 50 m thick (Fig. 2).

Calcareous rocks above them are represented by alternating beds of organogenic-detrital limestones of variable grain-size types. Clastic fractions of these rocks consist of rare foraminiferal tests and rounded fragments of echinoderms, corals, algae, pelecypods, and bryozoans. Inorganic clastic material of the rocks (5–20% of their volume) includes fine fragments of volcanic rocks, Fe-chlorite ooids, and grains of pelitomorphic limestones, quartz, and feldspar. According to their structure, cement type, and composition of clastic fraction, the rocks belong to the group of intrabiomicritic (occasionally sparry) packstones (grainstones) and rudstones. The limestone beds are 10 to 35 m thick in total. The first carbonate member is crowned with beds and lentils of volcaniclastic sandstone, gravel, and conglomerate, which are massive or show the indistinct horizontal, wavy, and cross laminations. Some beds exhibit fine rhythmic alternation (rhythms of 0.5 to 1 cm thick) of sandstone and siltstone laminae, material of which is well graded. Judging from the member structure, its rocks resemble coastal deposits of the "Crimean Oxfordian," a specific formation distinguished by Frolov (1998) and widespread in the lower Upper Jurassic of the Crimea, Cis- and Trans-Caucasian regions. Macro- and microscopic features of the rocks suggest that they accumulated within a very narrow and shallow sea zone adjacent to the coast (Voznesenskii *et al.*, 1998). Basal conglomerates of the member represent deposits of coastal setting with highly mobile hydrodynamics and of small deltaic fans, where they compensated the whole depth dropping. In the distal fan areas, they were replaced by facies of sandy to silty deposits with gravel and conglomerate lentils. The upper flattened parts of deltaic fans corresponded to accumulation settings of biogenic carbonate sediments. The carbonate clastic material descended in a form of aprons from these sites down to the deeper areas of delta slope and accumulated there in interlayers of biogenic-detrital sediments. A weak wave activity was favorable for the concurrent burial of rudaceous organic fragments and fine detritus of pelitic to aleuritic grade

that gave rise to accumulation of poorly sorted biogenic-detrital limestones with micritic matrix. Because of the transgressive trend of the Oxfordian basin evolution, the limestone beds of distal fan zones were buried under the deep-water sandy and gravely deposits (Fig. 3A).

Age of limestone beds in the first member is constrained by ammonites typical of the lower Oxfordian *Quenstedtoceras mariae* Zone, which have been found at the Chanakhchi and Antarashat localities (Gorbachik *et al.*, 1987). Foraminifers, which are unsatisfactory preserved, are rare at this level. Their species identified here are *Reophax* cf. *scorpiurus* Mont., *Belorussiella* sp., *Haplophragmoides* sp., *Nautiliculina* sp., and *Lenticulina* sp. (Plate 1).

The thick (1000–1300 m) volcanogenic–terrigenous sequence that overlies the basal carbonate member includes the extended lentils of organogenic-detrital limestones with sparite and micrite matrix. The lentils are up to 10–16 m thick, and limestones are the accumulation products of clastic material that have been produced by disintegration of small biogenic buildups in the coastal or nearby zone and reworked to a variable extent under influence of wave activity.

The foraminiferal assemblage that characterizes lentils of carbonate and terrigenous rocks includes the following taxa: *Glomospira gordialis* (Park. et Jones), *Glomospirella charoides* (Park et Jones), *Marssonella doneziana* Dain, *Ammobaculites verus* Dain, *A. hagni* Bhalla et Abbas, *Mayncina* sp., *Quinqueloculina mitchurini* Dain, *Sigmoilina microcostata* Danitch, *Nautiloculina* cf. *oolithica* Mohler, and *Trocholina alpina* (Leupold) (Plate 1). As the assemblage includes such forms as *Marssonella doneziana* Dain, *Quinqueloculina mitchurini* Dain, and *Sigmoilina microcostata* Danitch, the corresponding host deposits can be attributed to the Kimmeridgian Stage.

The second carbonate member is recognized in the middle part of the Upper Jurassic volcanogenic–sedimentary sequence that is composed of basic lavas, volcaniclastic sandstones, and conglomerates with thin lenticular intercalations of sandy organogenic-detrital limestones. The main carbonate interval of the member is 30–50 m thick, and rocks are represented here by gray to pinkish gray organogenic-detrital limestones. The biogenic clastic fraction of the latter includes fragments of corals, echinoderms, pelecypods, and occasional larger foraminifers. Grains of plagioclase and volcanic rocks occur in core parts of calcareous ooids from this fraction. Admixture of non-biogenic clastic grains, mostly of volcanic rocks, micritic limestones, and plagioclase, is insignificant, representing less than 10% of the rock volume. Matrix of the rocks usually consists of sparite cement that corresponds to the clean biogenic sand of a shallow sea zone with active hydrodynamics (intrabiosparite grainstones and rudstones). In rocks of the other type, which are less frequent, organogenic detritus is cemented by micritic mass that is an

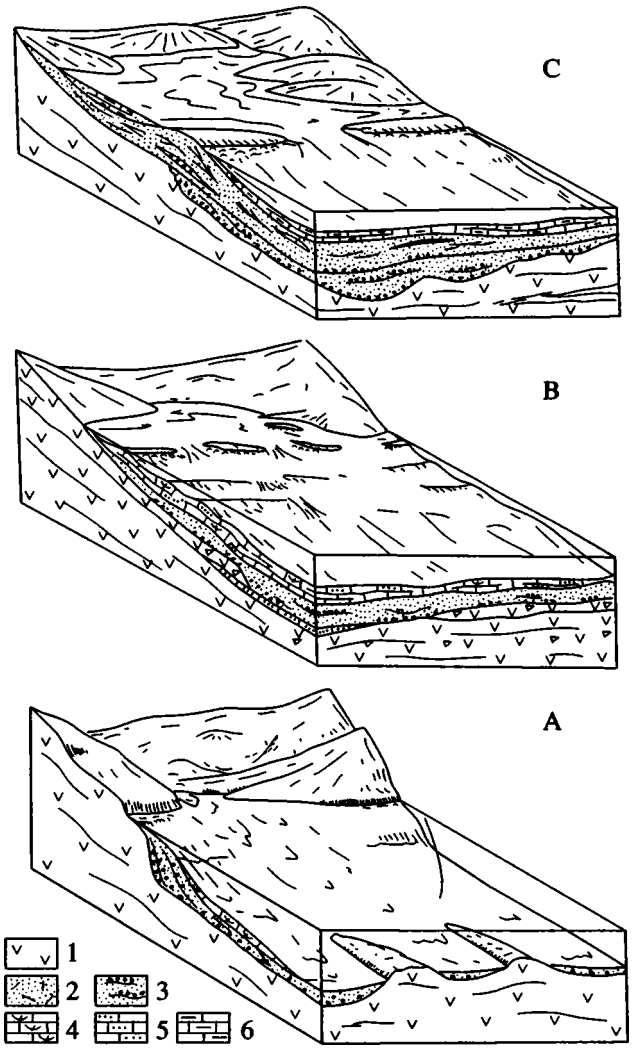
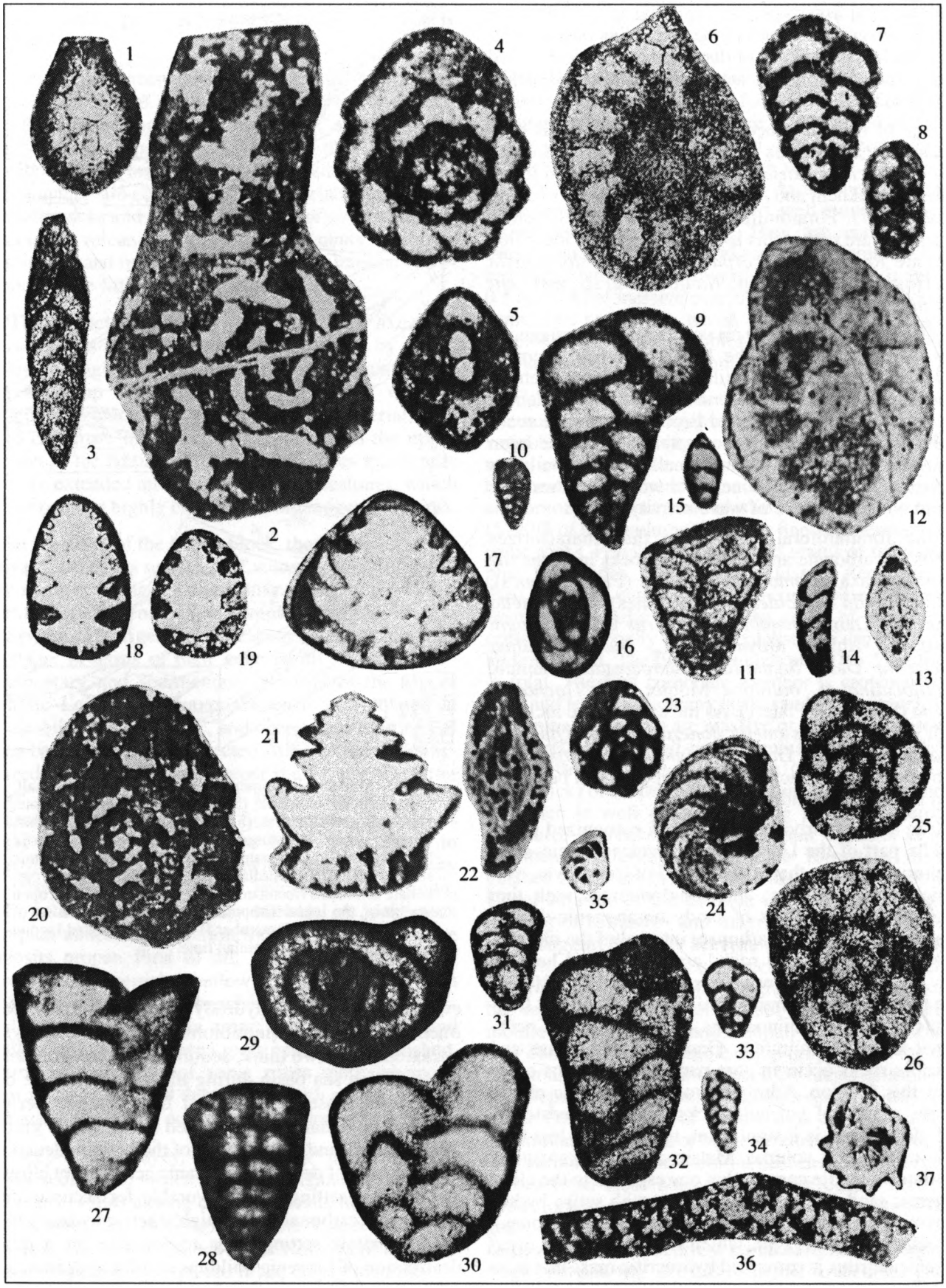


Fig. 3. Models of sedimentation: (1) andesite or basalt; (2) tuffaceous sandstone; (3) sandstone and conglomerate; (4) organogenic-detrital, (5) sandy, and (6) pelitomorphic, slightly clayey limestones. (A) Accumulation of clastic limy sediments (carbonate member I) in a coastal-deltaic setting, the initial Oxfordian time; (B) accumulation of carbonate sediments (member II) in a shallow zone of an open sea basin, the initial Tithonian time; (C) accumulation of carbonate sediments (member I) in a semi-isolated lagoon of a sea basin, the Valanginian time.

evidence of less active hydrodynamic conditions in sedimentation settings (intrabiomicritic wackestones or packstones). As we think, sedimentation environments in the Kafan sea basin during the Tithonian time differed from those of the initial Late Jurassic period, because the initially differentiated bottom relief gradually flattened, and coastal zone of the basin widened. In the moments of decreased volcanic activity, hot climate and shallow settings were favorable for accumulation of biogenic carbonates. The highly active hydrodynamics in shallow settings was responsible for a quick destruction of biogenic buildups and for appearance of

Plate I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offshore bars and mounds of well-graded carbonate sand (grainstones), which surrounded the accumulation areas of fine-grained clayey carbonate deposits (wackestones or packstones, Fig. 3B).

Carbonate rocks of the second member yield in abundance the diverse foraminifers. Assemblage of the latter includes the following taxa: *Saccammina* sp., *Everticyclammina* ex gr. *virguliana* Koechlin, *Pseudocyclammina lituus* (Yokoyama), *Melathrokerion spirialis* Gorbatchik, *Stomatostoecha* sp., *Charentia evoluta* Gorbatchik, *Textularia alexanderi* (Laliker), *Gaudryina bucowiensis* Cushman et Glazewski, *Lenticulina* sp., *Dentalina nana* Reuss, *Pseudonodosaria humilis* (Roemer), *Trocholina alpina* (Leupold), *T. elongata* (Leupold), and some other species (Plate 1). The majority of these taxa is known from the Tithonian deposits of the Tethyan belt, some of them, e.g., *Charentia evoluta* and *Pseudocyclammina lituus*, are characteristic of the Tithonian–Berriasian interval, and others are of a wider stratigraphic range (Kuznetsova and Gorbachik, 1985; Gorbachik *et al.*, 1987; Hottinger, 1967; Peybernes, 1976). In addition to sections of foraminiferal tests, we distinguished in thin sections the abundant remains of dasycladacean algae representing genera *Triplorella, Marcoporella*, and *Verticillorella*, which have been pictured and described by Jaffrezo (1973, 1974) who discovered them from the Jurassic–Cretaceous boundary deposits of France.

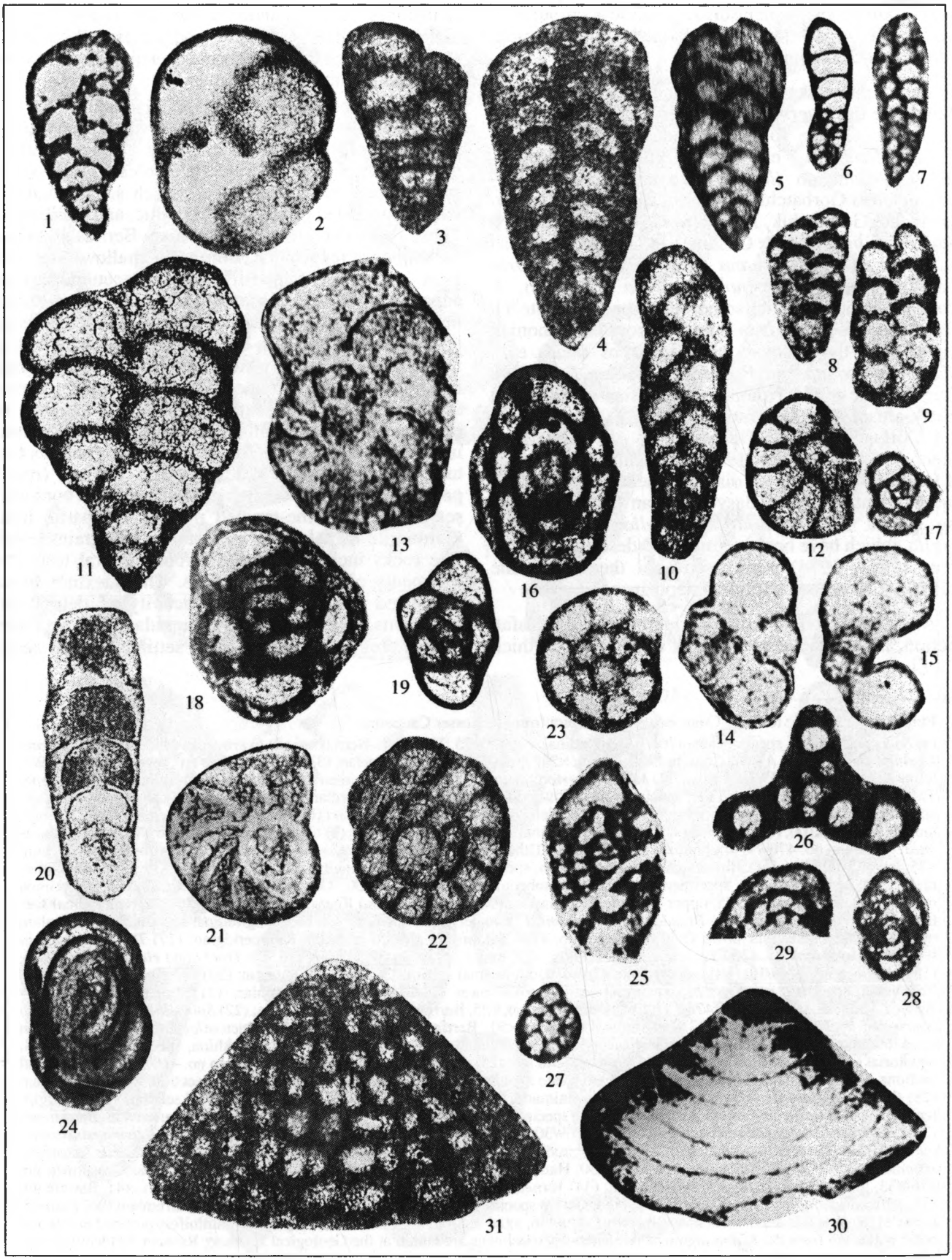
Above the second carbonate member of the Kafan section, a volcanic sequence that is 250–400 m thick includes separate lentils of tuffaceous and terrigenous sandstones and organogenic-detrital limestones. In their composition, structure, and genesis, these limestones are identical to those occurring below in the Oxfordian–Kimmeridgian interval.

The third carbonate member localized in the top portion of the Upper Jurassic interval includes basal beds of volcanogenic–sedimentary deposits (60 m thick), mostly of poorly graded volcaniclastic sandstones, conglomerate-breccias, which are intercalated with dolomitized psammitic, psephitic, and lapilli tuffs. These rocks overlying the Tithonian–Berriasian volcanic sequence represent a complex of shallow-water and even continental deposits, which accumulated in depressions of a rugged volcanic relief. Gray clayey limestones above those rocks enclose interlayers of poorly graded, massive, fine- to coarse- grained volcaniclastic sandstones, which are usually dolomitized. Limestones are thin- to medium-platty, frequently massive, dolomitized to a various extent, and showing an indistinct horizontal lamination and conchoidal fracturing. Under microscope, they reveal their grumous texture with aggregated fine biogenic detritus (round patches 0.1–1.0 mm in diameter with distinct contours) set in the very fine-grained micritic or sparite mass (pelmicrite to pelsparite packstones and grainstones). The rocks include occasional foraminiferal tests and fragments of small pelecypods. Their texture likely originated owing to the life activity of detrivorous organisms, when sediments accumulated, as we think, in sheltered lagoons and quiet settings behind sandy

Plate I. Late Jurassic to Early Cretaceous foraminifers from the Lesser Caucasus:

(1) *Sacccammina* sp., specimen no. 4769/1, longitudinal section, ×75, Tithonian–Berriasian; (2) *Everticyclammina* ex gr. *virguliana* Koehlin, specimen no. 4769/2, longitudinal section, ×23, upper Tithonian–Berriasian; (3) *Belorussiella ?* sp., specimen no. 4769/3, longitudinal section, ×75, Oxfordian; (4) *Melathrokerion spirialis* Gorbatchik, specimen no. 4769/4, equatorial section, ×100, upper Tithonian; (5) *Nautiloculina ?* sp., specimen no. 4769/5, axial section, ×45, Oxfordian–Kimmeridgian; (6) *Stomatostoecha* sp., specimen no. 4769/6, equatorial section, ×150, Tithonian; (7, 8) *Textularia alexanderi* (Lalicker): (7) specimen no. 4769/7, longitudinal section, ×100, (8) specimen no. 4769/8, longitudinal section, ×75, Tithonian; (9) *Gaudryina bucowiensis* Cushman et Glazewski, specimen no. 4769/9, longitudinal section, ×100, Tithonian; (10) *Textularia* sp., specimen no. 4769/10, longitudinal section, ×25, lower Tithonian; (11) *Marssonella doneziana* Dain, specimen no. 4769/11, longitudinal section, ×100, Kimmeridgian-Tithonian; (12) *Lenticulina* sp., specimen no. 4769/12; subequatorial section, ×150, Oxfordian; (13) *Lenticulina* sp., specimen no. 4769/13, axial section, ×25, upper Tithonian–Berriasian; (14) *Dentalina nana* Reuss, specimen no. 4769/14, longitudinal section, ×50, upper Tithonian; (15) *Pseudonodosaria humilis* (Roemer), specimen no. 4769/15, longitudinal section, ×50, Tithonian; (16) *Quinqueloculina mitchurini* Dain, specimen no. 4769/16, tangential section, ×100, Kimmeridgian; (17) *Trocholina alpina* (Leupold), specimen no. 4769/17, longitudinal section, ×25, upper Tithonian–Berriasian; (18, 19) *Trocholina elongata* (Leupold): (18) specimen no. 4769/18, (19) specimen no. 4769/19; longitudinal sections, ×25, lower Tithonian; (20) *Pseudocyclammina lituus* Yokoyama, specimen no. 4769/20, equatorial section, ×45, lower Berriasian–lower Valanginian; (21) *Trocholina companella* Arnaud-Vanneau, specimen no. 4769/21, longitudinal section, ×25, Berriasian–lower Valanginian; (22) *Spirophthalmidium inflatum* Antonova, specimen no. 4769/22, longitudinal section, ×50, Berriasian–Valanginian; (23) *Dacussiloculina ?* sp., specimen no. 4769/23, tangential section, ×45, Berriasian–lower Valanginian; (24) *Discorbis crimicus* Schokhina, specimen no. 4769.24, equatorial section, ×25, Berriasian; (25, 26) *Reticulinella* sp.: (25) specimen no. 4769/25, (26) specimen no. 4769/26, subequatorial sections, ×150, Valanginian; (27) *Dorothia praeoxycona* Moullade, specimen no. 4769/27, longitudinal section, ×150, Barremian; (28) *Pseudotextulariella salevensis* Charollais, Bronnimann, Zaninetti, specimen no. 4769/28, longitudinal section, ×75, Valanginian; (29) *Glomospira urgoniana* Arnaud-Vanneau, specimen no. 4769/29; subequatorial section, ×150, Hauterivian–Barremian; (30) *Gaudryina gradata* Berthelin, specimen no. 4769/30, longitudinal section, ×150, Barremian–Aptian; (31) *Bolivinopsis rhopaloidea* Arnaud-Vanneau, specimen no. 4769/31, longitudinal section, ×100, Barremian–Aptian; (32) *Gaudryina tuchensis* Antonova, specimen no. 4768/32, longitudinal section, ×150, Hauterivian–Barremian; (33) *Verneuilina* aff. *pharaonica* Said, specimen no. 4769/33, longitudinal section, ×45, Barremian; (34) *Verneuilina* sp., specimen no. 4769/34, longitudinal section, ×45, Barremian; (35, 37) *Lenticulina* sp.: (35) specimen no. 4769/35, 37) specimen no. 4769/37, equatorial sections, ×45, Barremian; (36) *Palorbitolina* cf. *lenticularis* Blumenbach, specimen no. 4769/36, axial section, ×35, Barremian–Aptian. Foraminifers pictured in this and other plates are from the Kafan region of Armenia. All specimens are stored at the Geological Institute, Russian Academy of Sciences, collection no. 4769.

Plate II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bars of a semiclosed coastal zone (Fig. 3C). Dolomitization of the whole member could be caused by basinal brines derived from thick underlying volcanics at the post-sedimentation stage. It is also possible that interstitial water concentrated magnesium at the stage of sedimentation under very arid climatic conditions, which were typical of the entire central segment of the northern Mesotethys periphery (Cis-Dobruja, Crimea, Caucasus, Central Asia) at the Jurassic–Cretaceous boundary time.

In addition to foraminifers occurring below, limestones of the third member yield the newly appearing forms *Trocholina molesta* Gorbatchik, *T. companella* Arnaud-Vanneau, *Spirophthalmidium inflatum* Antonova, and typical Valanginian species *Pseudotextulariella salevensis* Charollais, Bronnimann et Zaninetti and *Saracenaria valanginiana* Bartenstein et Brand (Plate 1; Arnaud-Vanneau, 1980; Gorbachik, 1971).

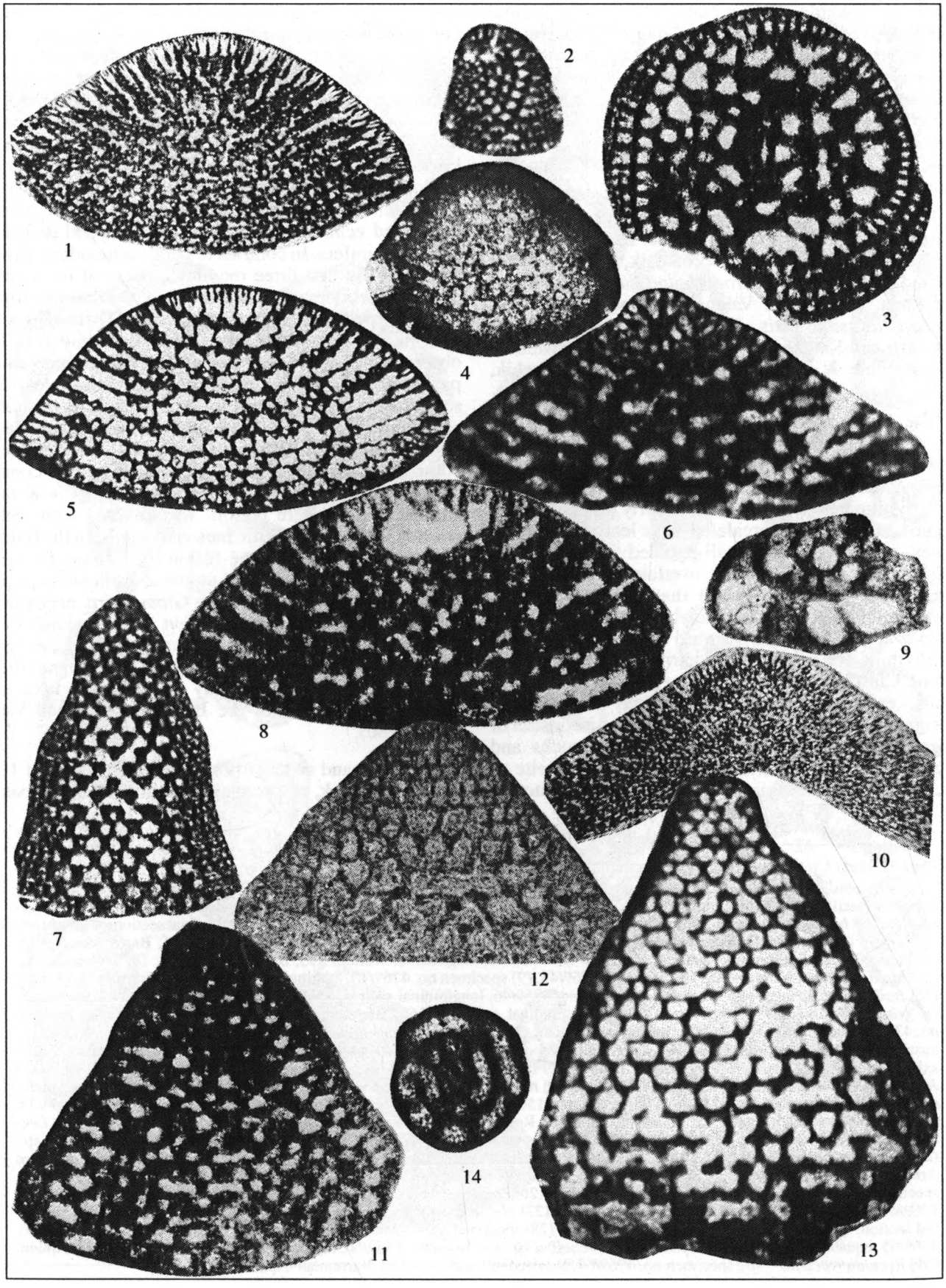
The fourth carbonate member spanning the Hauterivian–Barremian interval is clearly divisible into the basal terrigenous and greater calcareous parts. The terrigenous beds (30–40 m) are of a rhythmical structure. Each rhythm begins with beds of poorly graded coarse-grained sandstones intercalated with lentils of gravelstones and polymictic small-pebbled conglomerates (1.0–1.5 m). These rocks are overlain by well-graded fine-grained sandstone (1 m) that shows horizontal wavy lamination and occasional cross bedding with laminae angles decreasing toward the base. Terminal in the rhythms are organogenic-detrital limestones (0.6–1.0 m). Clastic fraction of the latter includes remains of corals, echinoderms, pelecypods, and foraminifers. Terrigenous admixture (10 to 25% in places) consists of plagioclase grains and clasts of volcanic rocks and micritic limestones. Cementing matrix is of sparite or micrite types (itrabiosparite grainstone or packstone). Below the base of some rhythms, there are erosion marks and incisions into underlying rocks. This rock complex presumably accumulated in deltaic fans of a shallow shelf or in contour-current and tide-wave channels of deeper slope settings. The upper part of the fourth member (250 m) is entirely composed of monotonous organogenic-detrital limestones of variable grain size. These limestones are pinkish gray, gray, and brown, massive in texture, often brecciated or cherty, having admixture of sandy and silty terrigenous material. Their clastic fraction includes remains of corals (rare) and echinoderms, abundant pelecypod detritus, and foraminifers. In contrast to coral–echinoderm limestones of the first three members, rocks of the fourth one are considerably richer in pelecypod detritus, foraminifers, pellets, and sponge spicules. Their matrix is micritic as a rule, though sparry cement can be observed as well (pelbiomicritic wackestones and packstones, biosparite grainstones and rudstones). In sandy and silty varieties, abundance of terrigenous admixture is up to 20% (volcanic clasts, plagioclase crystals, grains of micritic limestones). The described complex of sediments likely accumulated under conditions of high to low hydrodynamic activity in a vast shallow shelf, where bottom was covered with well-washed to clayey (micritic material) sands. In the fourth member, we identified the following foraminifer species: *Gaudryina tuchensis* Antonova, *Verneuilina pharaonica* Said, *Verneuilina* sp., *Glomospira urgoniana* Arnaud-Vanneau, *Melathrokerion nana* Arnaud-Vanneau, *M.* sp. *Lenticulina* sp., *Meandrospira* sp., *Bolivinopsis ropaloides* Arnaud-Vanneau, *B. ammovitrea* (Tappan) (Plates I, II). Many of the listed taxa are known beginning since the Barremian (Arnaud-Vanneau, 1980).

The fifth and sixth carbonate members (Fig. 2) 100 and 250 m thick, respectively, are situated in the basal

Plate II. Early Cretaceous foraminifers from the Lesser Caucasus:

(1, 2) *Arenobulimina cochleata* Arnaud-Vanneau: (1) specimen no. 4769/38, longitudinal section, ×45, (2) specimen no. 4769/39, transverse section, ×150, Barremian–Aptian; (3) *Gaudryina* sp., specimen no. 4769/40, longitudinal section, ×150, Aptian; (4, 5) *Verneuilinoides brevis* (Alekseeva): (4) specimen no. 4769/41, ×150, (5) specimen no. 4769/42, ×95, longitudinal sections; Barremian–Aptian; (6) *Gaudryinopsis minimus* Nikitina et Vassilenko, specimen no. 4769/43, longitudinal section, ×95, Barremian–Aptian; (7) *Verneuilinoides kaspiensis* Mjatliuk, specimen no. 4769/44, longitudinal section, ×95, Barremian–Aptian; (8, 9) *Bolivinopsis rhopaloides* Arnaud-Vanneau: (8) specimen no. 4769/45, (9) specimen no. 4769/45, longitudinal sections, ×70, Barremian–Aptian; (10) *Bolivinopsis ammovitrea* (Tappan), specimen no. 4769/46, longitudinal section, ×150, Barremian–Aptian; (11) *Gaudryina* ex gr. *gradata* Berthelin, specimen no. 4769/47, longitudinal section, ×150, Barremian–Aptian; (12) *Gavelinella* sp., specimen no. 4769/48, equatorial section, ×75, Barremian–Aptian; (13) *Gavelinella barremiana* Bettenstaedt, specimen no. 4769/49, equatorial section, ×120, Aptian; (14, 15) *Hedbergella* cf. *trocoidea* (Gandolti): (14) specimen no. 4769/50, (15) specimen no. 4769/51, axial sections, ×120, Aptian; (16) *Nummoloculina* sp., specimen no. 4769/52, tangential section, ×150, upper Barremian–lower Aptian; (17) *Rumanoloculina robusta* Neagu, specimen no. 4769/53, transverse section, ×45, Barremian–Aptian; (18) *Quinqueloculina* sp., specimen no. 4769/54, transverse section, ×150, Barremian–Aptian; (19) *Quinqueloculina* sp., specimen no. 4769/55, transverse section, ×45, Aptian; (20) *Nodosaria obscura* Reuss, specimen no. 4769/56, transverse section, ×150, Aptian; (21) *Lenticulina nikitinae* Vassilenko, specimen no. 4769/57, equatorial section, ×95, Barremian–Aptian; (22, 23) *Haplophragmoides* sp.: (22) specimen no. 4769/58, (23) specimen no. 4769/59, equatorial sections, ×150, Barremian–Aptian; (24) *Glomospira urgonica* Arnaud-Vanneau, specimen no. 4769/60, equatorial section, ×150, Barremian–Aptian; (25) *Mayncina bulgarica* Peybernes et Lang, specimen no. 4769/61, tangential section, ×45, Aptian; (26) *Quinqueloculina* aff. *lirellangulata* Loeblich et Tappan, specimen no. 4769/62, transverse section, ×150, Barremian–Aptian; (27) *Melathrokerion nana* Arnaud-Vanneau, specimen no. 4769/63, equatorial section, ×45, Aptian; (28, 29) *Melathrokerion* sp.: (28) specimen no. 4769/64, axial section (fragment), ×45, (29) specimen no. 4769/65, equatorial section, ×25, Aptian; (30) *Trocholina* sp., specimen no. 4769/66, longitudinal section, ×25, Barremian–Aptian; (31) *Rectodictyoconus* ? sp., specimen no. 4769/67, longitudinal section, ×150, Barremian–Aptian.

Plate III



(V) and uppermost (VI) intervals of the Aptian terrigenous–carbonate sequence. Both are composed of gray limestones very rich in sandy, silty, and clayey (up to the marl level) admixture. Clastic material represents 30–40% of the rock volume. Biogenic detritus of the rocks consists of pelecypod, foraminifer, coral, and echinoderm fragments. Somewhere, mostly near the top of the sixth member, rocks correspond to pure biomicrite and biosparite packstones and grainstones.

Carbonate deposits of the fifth and sixth members accumulated under conditions resembling those of the shallow sea basin of the Hauterivian–Barremian time (fourth member). It is though likely that the basin grew wider and was somewhat deeper later on.

In contrast to all other members described above, the fifth and sixth members are distinct because of presence of larger foraminifers of the order Orbitolinida. Among the latter, we identified *Palorbitolina lenticularis* Blumenbach, *Rectodictyoconus giganteus* Schroeder, *Falsurgonina pileola* Arnaud-Vanneau, *Iraqia simplex* Henson, *Paleodictyoconus arabicus* (Henson), and some others (Plates II, III). This assemblage of foraminifers is widespread in the Orgon facies of the Barremian–Aptian deposits in western Alps (Arnaud-Vanneau, 1980), Carpathian Mountains (Chernov, 1980), Turkmenistan (Mamontova, 1961), and other Tethyan regions.

At the level under discussion, orbitolinids coexist with *Glomospira urgonica* Arnaud-Vanneau, *Bolivinopsis rhopaloides* Arnaud-Vanneau, *Arenobulimina cochleata* Arnaud-Vanneau, *Verneuilinoides brevis* (Alekseeva), *Gaudryinopsis minimus* Nikitina et Vassilenko, *Gavelinella* sp., *Nummoloculina* sp., *Rumanoloculina robusta* Neagu, *Lenticulina nikitinae* Vassilenko, and *Trocholina* sp. In addition, the sixth member yields smaller foraminifers (Plate II), most interesting among which are *Gavelinella barremiana* Bettenstaedt and *Hedbergella trocoidea* (Gandolfi) (Plate II, figs. 13 and 14). The former one is of a subglobal distribution in the Barremian–Aptian deposits. The latter species represents a single planktonic form distinguished in the studied section. It is unknown in Barremian deposits and corresponds to the index species of the upper Aptian foraminiferal zone, if we accept the three-member zonation of this stage (Sigal, 1977; Gorbachik, 1986). Thus we may assume that the fifth member is of the Barremian–lower Aptian age, and the sixth member should be attributed to the middle–upper Aptian.

CONCLUSION

The almost identical structure and composition of coeval volcanogenic–sedimentary deposits in the Kafan, Karabakh, and Agdam zones of the Lesser Caucasus allow us to share the formerly suggested idea (Knipper, 1975; Sokolov, 1977) that the Jurassic–Early Cretaceous sequence of the Kafan region accumulated within a volcanic arc that was located in the northern active margin of the Tethys.

The Late Jurassic–Early Cretaceous sedimentation in the Kafan region can be divided in two stages. The first one developed in the Oxfordian–Berriasian period, when a very rugged volcanic relief was formed, and the corresponding sea basin had a narrow steep shelf with the high depth gradients between the coastal and open-sea zones. This stage was distinguished by progressively growing abundance of small bioherms, which produced wide aprons of biogenic clastic material against the background of dominant pyroclastic sedimentation and deposition of volcanogenic–sedimentary material in shallow-water settings.

The next stage that begun since the Valanginian time corresponded to the period of considerably decreased volcanic activity. Simultaneously, it was a time of active denudation and peneplanation, when sea basin grew wider so that its shelf zone became wide and gently inclined. Ingression episodes at the beginning of the second stage and the accelerated supply of terrigenous material were favorable for appearance of lagoonal sedimentation settings in the coastal sea zone. The widespread development of bioherms in the first half of the Early Cretaceous and their aprons of clastic material gave rise to accumulation of carbonate sands in many areas of the shallow zone, where tidal cycles and wave activity led to formation of sand bars, submarine hills and ridges. Intensity of biogenic sedimentation somewhat decreased during the initial Aptian time. Though a weak volcanic activity was still in progress during the second stage of the region development, the terrigenous and biogenic sedimentation dominated at that time.

Plate III. Early Cretaceous foraminifers from the Lesser Caucasus:

(1) *Palorbitolina* sp., specimen no. 4769/68, axial section, ×25, Barremian–Aptian.; (2) *Paleodictyoconus arabicus* (Henson), specimen no. 4769/69, axial section, ×25, Aptian; (3) *Cribellopsis* ? sp., specimen no. 4769/70; oblique transverse section, ×65; Aptian; (4) *Paleodictyoconus* sp., specimen no. 4769/71, axial section, ×45, Aptian; (5) *Paleodictyoconus* sp., specimen no. 4769/72, axial section, ×35, Barremian–Aptian; (6) *Falsurgonina pileola* Arnaud-Vanneau et Argot, specimen no. 4769/73, axial section, ×65; Barremian–Aptian; (7) *Iraqia simplex* Heson, specimen no. 4769/74, axial section, ×75, Barremian–Aptian; (8) *Palorbitolina lenticularis* Blumenbach, specimen no. 4769/75, axial section, ×65; Barremian–Aptian; (9) *Falsurgonina pileola* Arnaud-Vanneau et Argot, specimen no. 4769/76, tangential section in the embryonic part of megalospheric test, ×150, Barremian–Aptian; (10) *Dictyorbitolina* sp., specimen no. 4769/77, axial section, ×45, Barremian–Aptian; (11) *Paleodictyoconus arabicus* (Henson), specimen no. 4769/78, axial section, ×65; Barremian–Aptian; (12) *Rectodictyoconus* sp., specimen no. 4769/79, axial section, ×75, Barremian–Aptian; (13) *Rectodictyoconus giganteus* Schroeder, specimen no. 4769/80, axial section, ×45, Barremian–Aptian; (14) *Meandrospira* sp., specimen no. 4769/81, equatorial section, ×150, Hauterivian–Barremian.

The analyzed changes in foraminiferal assemblages from the studied sediments show that their evolution is also divisible in two periods connected with different stages of the basin development. The first period of Oxfordian–Tithonian to initial Berriasian time is characterized by paleocoenoses of predominantly agglutinated foraminifers representing orders Ammodiscida Lituolida, Ataxophragmiida, and Textulariida (genera *Glomospira, Glomospirella, Ammobaculites, Mayncina, Marssonella*, and *Textularia*). All foraminifers that inhabited basin during this period represented a benthic community, in which secretory forms were those of Miliolida and Involutinida orders (genera *Qninqueloculina, Sigmoilina*, and *Trocholina*). All the taxa listed above are of a wide geographic range.

By the end of the first period, during the Tithonian–Berriasian time, the taxonomic composition of foraminiferal community changed because of appearance of endemic Tethyan genera. These were genera *Everticyclammina, Pseudocyclammina, Melathrokerion, Stomatostoecha*, and *Charentia* (order Lituolida), the characteristic dwellers of warm and shallow sea zone. As before, secretory forms were of subordinate significance, and newcomers among them represented order Lagenida (genera *Dentalina, Lenticulina*, and *Pseudonodosaria*). All these genera were cosmopolitan.

Since the Valanginian time that marks the commencement of next stage in development of biota and sedimentation in the basin, the foraminiferal community became impoverished and experienced next renewal. This moment marks appearance of genera *Reticulinella* (order Lituolida) and *Pseudotextulariella* (order Ataxophragmiida).

An important step in the further evolution of foraminiferal community was connected with appearance of orbitolinids, the Tethyan endemics whose skeletons are of a heterogeneous inner structure, and which became distinctly dominant during the Barremian–Aptian time. These were genera *Palorbitolina, Rectodyctioconus, Falsurgonina, Iraqia, Paleodyctioconus*, and *Dictyorbitolina*, the habitants of shallow tropical basins. Another important evolutionary event corresponded to appearance of first and sparse planktonic forms representing genus *Hedbergella* (order Globigerinida).

Thus, the foraminiferal community evolved in parallel with changes in the character of sedimentation and reflected corresponding changes in the habitat environments. On the other hand, the community developed depending not on the external factors only, but also on the internal biological properties of foraminifers and on the general trend of their evolution.

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The Global Upper Cretaceous Scale

A. G. Olfer'ev* and A. S. Alekseev**

**State Unitary Enterprise Geosintez-Tsentr, Varshavskoe sh. 39a, Moscow, 113105 Russia*

***Moscow State University, Vorob'evy gory, Moscow, 119899 Russia*

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Abstract—The global stratigraphic scale of the Upper Cretaceous biozonation is considered. The presented version is based on stage boundaries of the Cretaceous System proposed by the International working groups and discussed at the Second International Symposium in Brussels (1995). For maximum convenience, the scale includes several zonations of different orthostratigraphic faunal groups: ammonites, planktonic foraminifers, and calcareous nannoplankton. In addition, the scale is supplemented with the magnetostratigraphic scheme and radiological dates of stage boundaries. The scale can serve as a reliable tool for correlation of the Cretaceous regional subdivisions in different regions of Russia.

Key words: Cretaceous System, Upper Cretaceous deposits, stage, zone, ammonites, planktonic foraminifers, calcareous nannoplankton.

INTRODUCTION

As is known, stage is a basic unit of the international chronostratigraphic scale. This seems to be permissible for the global scale because it would be difficult to directly operate with zonal units and to correlate them. Nevertheless, the main destination of any chronozone is to fix some points at the time vector. These points should form the internal framework of a stage. The chronozone is an abstract unit, which does not necessary enclose the taxon, after which it is named.

The intense activity of the International Commission on Stratigraphy during last two decades that was aimed to define the zonal unit boundaries in the global scale required new tools enabling the more detailed stratigraphic subdivision. Although the stage boundary is usually defined at the appearance level of a single taxon, selection and correlation of boundary intervals are performed in fact with the help of zonal units. This is to a certain extent admitted by new leaders of the International Commission on Stratigraphy (Gradstein, 2000). In addition, practical activity, particularly in relatively small areas, requires a stratigraphic tool with resolution higher than a stage or substage, and necessary accuracy is characteristic of a chronozone. Therefore, the requirement of the Russian Stratigraphic Code to include also zones in the global scale is rightful.

Many Russian researchers share a standpoint that zones are mainly of the correlative significance and substantiate a special importance of the so-called "biozonal standards" (Koren' *et al.*, 2000). These units, in fact, perform a function of chronozones of a global scale, but are not included into the latter, and such a situation is not fortunate. The idea to include into the global scale the units subordinate to the stage accords also with a general tendency to detail the stratigraphic schemes and to achieve their higher resolution (Gladenkov, 2001).

In Russia, most difficult situation is with elaboration of the global Upper Cretaceous scale lacking approval until now that substantially impedes progress in stratigraphic research and geological mapping. Recent significant achievements in the Upper Cretaceous stratigraphy of West Europe and North America, as well as the analysis of available data on the Upper Cretaceous sections of the Russian platform, open possibility to propose such a scale characterized in this work (Table 1). Problems related to elaboration of the regional Upper Cretaceous scheme for the Russian platform will be considered in a special publication.

STAGE SUBDIVISIONS

The global Upper Cretaceous scale is yet far from the eventual approval (Remane *et al.*, 2000). At present, only the upper boundary of the Cretaceous System is officially approved to be coincident with the base of the Danian Stage. At the same time, if scales of some other systems lack definiteness even in a number of stages, not only in their names, stage subdivision of the Cretaceous System is widely accepted and nobody is going to change it. It is necessary only to select the appropriate and well-correlated levels that could serve as boundaries of traditional stages and to find the unambiguous Global Stratotype Sections and Points (GSSP) for their limitotypes. This means that the Upper Cretaceous scale of Russia can retain the traditional subdivision into West European stages. The proposed version of the scale is mainly based on boundaries of Cretaceous stages, which have been suggested for consider-

ation by the International working groups at the Second International Symposium held in Brussels, September 8–16, 1995, and dedicated precisely to this problem (Tröger and Kennedy, 1996; Bengston, 1996; Kauffman *et al.*, 1996; Lamolda and Hancock, 1996; Hancock and Gale, 1996; Odin, 1996).

PRINCIPLES OF CONSTRUCTION

The global scale and its zonal subdivision should be at maximum convenient for using. This means that the scale should include several, not one, successions of zones, because it is difficult to choose one zonation as a universal at the modern level of knowledge.

After pioneering investigations by A.P. Pavlov and A.D. Arkhangel'skii, the Russian platform turned into the birth region of domestic concepts concerning the Upper Cretaceous stages. Unfortunately, ammonites, the leading group of fossils in the Cretaceous stratigraphy, are extremely rare in the Russian platform. Accordingly, the other groups of macrofauna, e.g., inocerams, belemnites (mainly), and some others, were used to elaborate the regional stratigraphic zonations (Naidin, 1960, 1969; Gerasimov *et al.*, 1962; Naidin *et al.*, 1984). Inasmuch as the established zonal subdivision do not correlate precisely with West European ammonite- and even belemnite-based standard units, it would be unreasonable to select them for basic units of a global scale.

The proposed scale includes a succession of ammonite zones established in West Europe. This succession has an important function to determine the stages ranges, which also were defined first in Europe. The direct application of the ammonite scale is difficult, however, because ammonites are absent in Cretaceous sections of the Russian platform, extremely rare in West Siberia, and have endemic character in the Far East.

Therefore, the scale includes zonal subdivisions based on planktonic foraminifers and calcareous nannoplankton. The planktonic foraminifer zones are taken from the last version of the Mediterranean (Tethyan) scale (Robaszynski and Caron, 1995; Premoli Silva and Sliter, 1995; Hart, 1999), in which they are reliably correlated with boundaries of magnetostratigraphic units, though not always with the units in the standard ammonite scale. The Crimean–Caucasian scale by Maslakova (1978) that is close in some intervals to the Tethyan one is disadvantageous in the last respect. It should be noted as well that the accepted planktonic foraminifer zonation can be traced in the North Caucasus, Ciscaucasia, and North Caspian depression.

Calcareous nannoplankton is a highly promising practical tool for age determination of local and regional stratigraphic units in the Russian platform and North Caucasus. The proposed scale includes two nannoplankton zonations elaborated for the Mediterranean (Perch-Nielsen, 1985) and oceanic (Roth, 1978) regions. The combination of both, principally close zonal scale allows subdivision of the entire stratigraphic interval under consideration and enables a sufficiently reliable correlation with planktonic foraminifer zonation.

As was recently established, stratigraphic ranges of some index species of calcareous nannoplankton in middle latitudes of the northern hemisphere, where the territory of Russia is located, differ noticeably from those in the Mediterranean region and tropical areas of the oceanic realm (Hardenbol *et al.*, 1998). This complicates the situation, but in view of insufficient knowledge of nannofossils in northern regions, the difficulty seems overestimated.

The potential of calcareous nannoplankton studied in West Siberia, particularly in its southern part, and in the Far East is inadequately used. In the later region, the Upper Cretaceous reference section in the Naiba area of Southern Sakhalin and sections of Western Kamchatka are of particular importance and need additional study. These sections are characterized by ammonites and were examined earlier by Pergament (1974).

In future, the scale can be supplemented with the radiolarian zonation that can help to solve stratigraphic problems in northern Russia and in areas with widespread deep-water siliceous sequences, e.g., in Sakhalin, Kamchatka, and in the Koryak Highland.

It should be emphasized that some other fossil groups such as inocerams and echinoderms have also a high stratigraphic potential, as is justly noted by Naidin (1996). Moreover, the International working groups on the Coniacian, Santonian, and Campanian stages suggested to place the boundary between first two stages at the appearance level of some inoceram-like mollusks, and the lower boundary of the Campanian Stage at the extinction level of marine crinoid genus *Marsupites*. The same levels can reliably be recognized in the West European ammonite zonation, as it is shown in the next section.

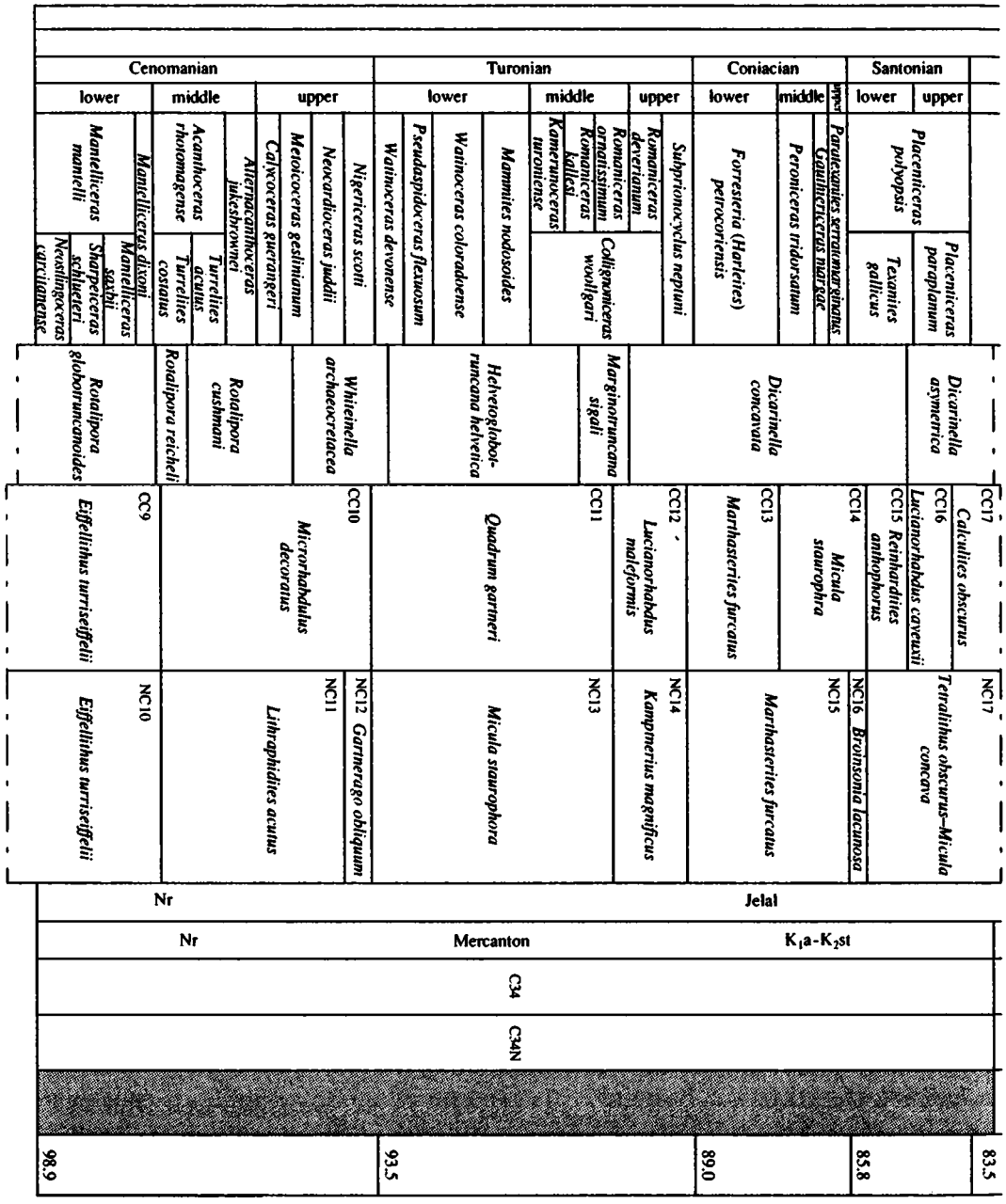
STAGE BOUNDARIES AND AMMONITE ZONATION

Cenomanian

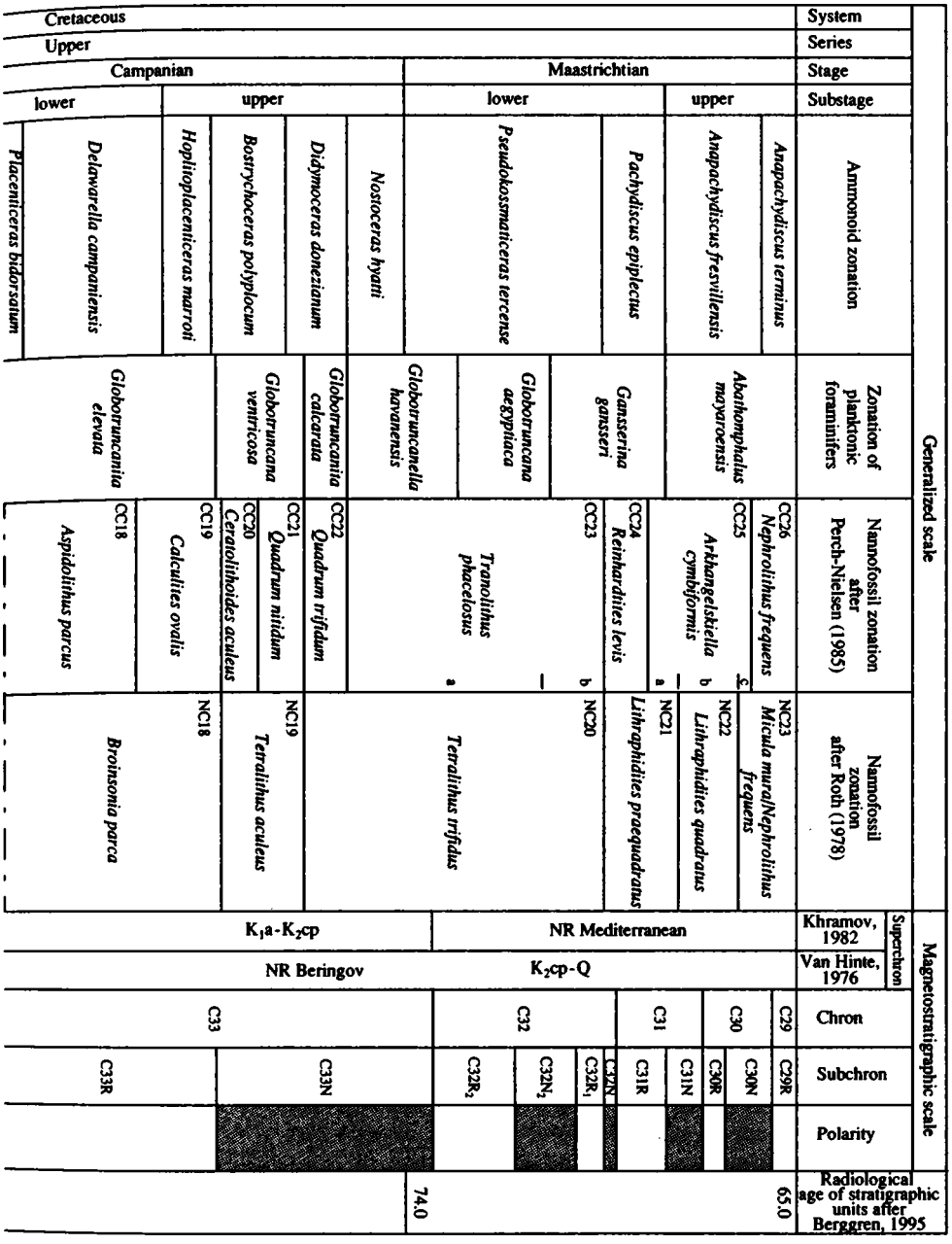
This stage is traditionally subdivided into three substages. The International working group (Tröger and Kennedy, 1996) selected the planktonic foraminifer species *Rotalipora globotruncanoides* Sigal as the index species for the lower boundary of the stage and recommended to establish the limitotype (GSSP) of the Albian–Cenomanian boundary in the Mont Risou section of Provence, France. This level is located 6 m below the first appearance level of ammonite *Neostlingoceras carcitanense* (Matheron) that marks the base of the Cenomanian Stage in Europe.

In the general scale proposed here, the base of the *Mantelliceras mantelli* Zone traditional for the type locality is selected to mark the lower boundary of the Cenomanian Stage. In line with suggestion of

Table 1. The global Upper Cretaceous stratigraphic scale



W.J. Kennedy (Gale *et al.*, 1996), this zone is subdivided into three, not two, parts: (from the base upward): *Neostlingoceras carcitanense, Sharpeiceras schlueteri,* and *Mantelliceras saxbii* subzones. The lower–middle Cenomanian boundary is traditionally accepted at the level of the replacement of beds with *Mantelliceras* by layers with first representatives of the *Acanthoceras* and *Cunningtoniceras* genera. In the proposed scale, this level corresponds to the boundary between the *Mantelliceras dixoni* and *Acanthoceras rhotomagense* zones, although, in opinion of Atabekyan (1994), *Cunningtoniceras cunningtoni* (Sharpe) appears earlier than *Acanthoceras rhotomagense* (Brgt.). Kennedy (1995) believes that both ammonite species appear synchronously. The limitotype of the middle Cenomanian Substage is selected in the section exposed in the Southerham Grey Quarry of Lewes (Sussex County, England), where the GSSP is established at the first appearance level of ammonite species *Cunningtoniceras interme* (Perv.). Inasmuch as there is no unambiguous view on position of the middle–upper Cenomanian boundary, it is placed in the proposed scale at the same level as in the stratotype region, i.e., between the *Alternacanthoceras jukesbrownei* and *Calycoceras guerangeri* zones (Wright *et al.*, 1984), although some specialists suggest to draw it either at the base of the



first of these zones or at the top of the second one (Tröger and Kennedy, 1996).

Turonian Stage

The Cenomanian–Turonian boundary position experienced most significant changes during last years. In accordance with resolutions approved at the colloquia of 1976 and 1981 (Robaszynski, 1983; Atabekyan *et al.*, 1986) and subsequently at the First International Symposium on Cretaceous Stage Boundaries in Copenhagen (Hancock, 1984; Kennedy, 1984), it was placed in West Europe (England, northern France) between the *Neocardioceras juddii* and *Watinoceras coloradoense* zones. However, some researchers assumed a significant hiatus at this level in Europe (Robaszynski and Amedro, 1980; Robaszynski, 1983), and later on its existence was confirmed. Therefore the section of the Rock Canyon anticline in the Arkansas River basin west of Pueblo, Colorado, United States, was suggested to be the GSSP of the Turonian Stage base. Kennedy and Cobban (1991) defined here the *Nigericeras scotti* Zone, which is located above the *Neocardioceras juddii* Zone and terminates the Cenomanian sequence, and the overlying *Watinoceras devonense, Pseudoaspidoceras flexuosum*, and *Vascoceras birchbyi* zones in the lower-

most Turonian Stage. Recently, this section was revised by I. Walaszczyk (Kennedy *et al.*, 2000), and, as a result, the *Watinoceras devonense* and *Pseudoaspidoceras flexuosum* zones were included into the basal part of the lower Turonian (Bengtson, 1996). Above the last zone, the *Watinoceras coloradoense* Zone was placed instead of the *Vascoceras birchbyi* Zone distinguished by Kennedy and Cobban, because *Watinoceras coloradoense* (Hend.) appears in Bed 97 of the Rock Canyon section simultaneously with *Vascoceras birchbyi* Cobban et Scott (Kennedy and Cobban, 1991). The idea to place the lower boundary of the Turonian Stage at the base of the *Watinoceras devonense* Zone of the Rock Canyon section was supported by most of members of the International Working Group and forwarded for approval to the Subcommission on the Cretaceous System. In the proposed general scale, that boundary is placed exactly the this level.

The overlying part of the Turonian Stage is subdivided into zones in line with recommendations of the colloquium held in Turon (Robaszynski, 1983). The lower boundary of the middle Turonian corresponds as before to the first appearance level of ammonite *Collingnoniceras woollgari* (Mant.) found in Bed 120 of the Pueblo section. As the limitotype of the lower boundary of the upper Turonian has not been proposed by the International Working Group, the middle–upper Turonian boundary is placed between the *Romaniceras ornatissimum* and *Romaniceras deverianum* zones in line with recommendations of the Turon colloquium (Bengtson, 1996).

Coniacian Stage

The International Working Group on the Coniacian Stage recommended to place its lower boundary at the first appearance level of the inoceram species *Cremnoceramus rotundatus* (sensu Tröger) and to consider the base of Bed MK-47 in the Salzgitter-Salder quarry near Hannover, Lower Saxony (Germany), as representing the proper limitotype. This bed marks the ecological event *Didymotis* II responsible for mass appearance of *Cremnoceramus waltersdorfensis waltersdorfensis* (Andert). It also coincides with the extinction level of the ammonite species *Prionocyclus germari* (Reuss) characteristic of an upper part of the terminal Turonian zone and with first occurrence of *Forresteria petrocoriensis* (Coq.) (Kauffman *et al.*, 1996).

It is proposed to divide the Coniacian Stage into three substages. Although the GSSPs for the substage boundaries are not yet selected, the substage ranges are established rather reliably. The lower boundary of the middle substage is determined by the first appearance of the inoceram species *Volviceramus koeneni* (Möll.) and ammonite *Peroniceras tridorsatum* (Schlüt.). The base of the upper substage corresponds to the appearance level of the inoceram species *Magadiceramus subquadratus* (Schlüt.) (Kauffman *et al.* 1996). In line with conclusions of Marcinowski and Walaszczyk (Marcinowski *et al.*, 1996), the upper Coniacian Substage is accepted as corresponding to the *Paratexanites serratomarginatus* ammonite zone, and the *Cauthericeras margae* Zone is included, contrary to the opinion by Kennedy (1984), into the middle Coniacian Substage.

Santonian Stage

Because this stratigraphic interval is insufficiently studied, the International Working Group did not arrived at a definite conclusion with respect to the GSSP of the Santonian Stage base and recommended by the majority of votes to consider the first appearance level of the inoceram species *Cladoceramus undulatoplicatus* (Roem.) as marking that base. The First Symposium on the Cretaceous Stage Boundaries acknowledged this level, in addition to that of the first appearance of the ammonite subgenus *Texanites* (*Texanites*), as the best one for defining the Coniacian–Santonian boundary. Moreover, the appearance of both taxa was considered to be virtually synchronous (Birkelund *et al.*, 1984). Recently, it was established, however, that in some regions (Austria, Germany, northern Spain, Texas in the United States), ammonites of the *Texanites* (*Texanites*) subgenus appear along with the boreal inoceram species *Sphenoceramus cardissoides* (Goldf.) and *S. pachti* (Arkh.) earlier than *Cladoceramus undulatoplicatus* (Roem.). Based on results obtained in the Mönster Basin of North Westphalia (Germany), Kaplan and Kennedy (2000) recommended therefore to transfer the *Texanites pseudotexanus* (=*Texanites texanus*) ammonite zone and corresponding *Sphenoceramus cardissoides–S. pachti* inoceram zone into the Coniacian Stage as terminal units of the latter and to begin the Santonian section with the *Cladoceramus undulatoplicatus* inoceram zone and synchronous *Kitchinites emshericus* ammonite zone (Table 2). In our opinion, this variant is premature at least, because first representatives of *Cladoceramus undulatoplicatus* (Roem.) appear in England below the occurrence level of *Sphenoceramus cardissoides* (Goldf.) and *S. pachti* (Arkh.). In France and Spain, beds with *Magadiceramus subquadratus* (Schlüt.) are immediately overlain by deposits yielding *Cladoceramus undulatoplicatus* (Roem.), whereas in Austria (Mölbach section) *Cladoceramus undulatoplicatus* (Roem.) and *Texanites quinquenodosus* (Redten.) occur simultaneously (Tröger and Summesberger, 1994). These uncertainties should be clarified in the future, and now the Santonian Stage is accepted in our scale, following Hancock (1991), as corresponding to the *Placenticeras polyopsis* zone. We retained as well its two-member subdivision into *Texanites gallicus* and *Placenticeras paraplanum* zones, although the International Working Group considers the three-member subdivision of the Santonian as more preferable.

Table 2. Zonal subdivision of Santonian deposits in Europe (modified after Kaplan and Kennedy, 2000)

Kaplan and Kennedy, 2000			Seitz, 1961; Tröger, 1989	Hancock and Gale, 1996	Schulz *et al.*, 1984	Olfer'ev and Alekseev, this work			Olfer'ev and Alekseev, in press
Stage	substage	Ammonite zones of the Münster basin	Inoceram zones	Crinoid zones	Biozones of the Lägerdorf section (Germany)	Project of the global stratigraphic scale			Biozones of the East European platform
						Stage	substage	Ammonite zone, subzone	
Campanian	lower	*Placenticeras bidorsatum*			*lingua/quadrata granulata–quadrata*	Campanian	lower	*Placenticeras bidorsatum*	*Belemnitella praecursor mucronatiformis*
Santonian	upper	*Boehmoceras arculus*	*Sphenoceramus patootensiformis* (zone 29)	*Marsupites testudinarius*	*testudinarius/granulata*	Santonian	upper	*Placenticeras paraplanum*	*Sphenoceramus patootensis/Belemnitella praecursor praepraecursor*
				Uintacrinus socialis	*socialis/granulata*				
	middle	*Kitchinites emscheris*	*Sphenoceramus pinniformis* (zone 28)		*rogale/westfalica–granulata*		lower	*Placenticeras polyopsis*	*Texanites texanus/Sphenoceramus cardissoides/ Belemnitella propinqua propinqua*
					rogale/westfalica				
			Cordiceramus cordiformis (zone 27)		*coranguinum/westfalica*			*Texanites gallicus*	
	lower		*Cladoceramus undulatoplicatus* (zone 26)		*pachti–undulatoplicatus*				
Coniacian	upper	*Texanites pseudotexanus*	*Sphenoceramus pachti* (zone 25)						
		Paratexanites serratomarginatus	*Magadiceramus subquadratus* (zone 24)		*bucailli/praewestfalica*	Coniacian	upper	*Paratexanites serratomarginatus*	*Magadiceramus subquadratus*

Campanian Stage

The International Working Group on the Campanian Stage did not formulate eventually its conclusion with respect to the base of this stratigraphic unit, although there was a preliminary agreement that it should coincide with the top of the *Marsupites testudinarius* Zone. This zone should correspond, in principle, to the base of the *Placenticeras bidorsatum* Zone, as it was proposed by de Grossouvre (1901). In line with viewpoint of this researcher, the Campanian Stage was subdivided until recently into two substages spanning the *Placenticeras bidorsatum, Delawarella campaniensis, Hoplitoplacenticeras marrotti*, and *Bostrychoceras polyplocum* zones, although some of significant hiatuses have been established in the Campanian sections of Europe (Hancock, 1991; Burnett, 1998; and others). We retained these units in the proposed scale. According to data by Kennedy and Cobban (1999), the stage is most complete in North America, where they defined 22 local ammonite zones in the Pacific coastal sections, correlation of which with the zones of the Aquitanian stratotype region is unclear. In addition, taking into account the idea to fix the GSSP of the Maastrichtian lower boundary in the Tercis section (southwestern France) and to place, accordingly, the Campanian–Maastrichtian boundary significantly lower than now (Blaszkiewicz, 1980; Hancock *et al.*, 1993), we included two more zones (*Didymoceras donezianum* and *Nostoceras hyatti*) into the Campanian interval of our scale. In future, they may be regarded as zones of the third, upper substage.

Maastrichtian Stage

Biozonation and positioning of the lower boundary of the Maastrichtian Stage appear to be the complicated problems. At the First Symposium in Copenhagen, several level were considered as marking the base of this stratigraphic unit:

(a) first appearance level of ammonite species *Hoploscaphites constrictus* (J. Sow.);

(b) first appearance level of ammonite species *Pachydiscus neubergicus* (von Hauer);

(c) base of the *Belemnella lanceolata* belemnite zone;

(d) base of the *Globotruncana falsostuarti* Zone of planktonic foraminifers.

In addition, it was proposed to place the lower Maastrichtian boundary at the disappearance level of the planktonic foraminifers *Globotruncanita calcarata* (Cushm.) or at that of calcareous nannofossils *Quadrum trifidum* (Stradn.) Prins et Perch-Nielsen. One more idea of the last decade was to define the lower boundary of the Maastrichtian Stage at the appearance or, to the contrary, disappearance level of ammonites *Nostoceras hyatti* (Steph.) and at the extinction level of nannofossils *Aspidolithus parcus constrictus* (Stradn.) Hatt. (Odin, 1996).

Regardless any potential stratigraphic level that can be chosen for the GSSP of the Maastrichtian Stage base, it is clear that the Campanian and Maastrichtian stratotype sections are inappropriate for establishing the limitotype. In classical sections of the Charante Department in Aquitaine (northwestern France) that is a stratotype region of the Campanian Stage, the Maastrichtian has not been established so far (Atabekyan *et al.*, 1986). In the stratotype section exposed in the ENCI Quarry near Maastricht (Limburg Province in the southern Netherlands), the Maastrichtian Stage is represented only by its uppermost part overlying, with a significant hiatus, preserved fragments of the Campanian strata. "Tufaceous Maastricht Chalk" or the Maastricht Formation corresponds here only to the upper substage (*Belemnitella junior* and *Neobelemnella kazimiroviensis* zones) in the general scale of Russia. Thin underlying beds of "Gulpen Chalk" (Gulpen Formation) also belong mostly to the upper Maastrichtian, and only the lowermost layers yield early Maastrichtian belemnites of the *Belemnella occidentalis* (=*Belemnella sumensis*) Zone. Other artificial and natural exposures, which were considered as candidates for the GSSP selection, are the Boreal section of Kronsmoor Quarry in the lower course of the Elba River, northern Germany, and a series of Tethyan sections. The latter are situated in the Basque Zumaya Province of Spain, in the Bidart locality of France, and also in Tunisia, the Apennines (Italy), and southwestern France (Tercis section in the Landes). Most appropriate among all these sections is the last one, because it yields all main groups of macrofauna (except for belemnites), which can be used for correlation with the Boreal counterparts. The preservation state of foraminifers and nannofossils in this section is worse than in deep-sea sediments, but they are determinable nevertheless. As a result of voting on October 15, 1995, i.e., one month after the Brussels symposium, the Tercis section was recommended for the limitotype of the Maastrichtian lower boundary. It was proposed to be at the base of Member N_3, close to the first appearance level of *Pachydiscus neubergicus* (von Hauer). Hancock *et al.* (1993) who studied the Tercis succession, placed this boundary between the *Nostoceras hyatti* and *Pachydiscus epiplectus* zones. However, the index species of the first zone was not found in Members M, N_1, and N_2 of the Tercis section, which were included into the *Nostoceras hyatti* (s.l.) Zone by mentioned authors. This fact allowed P.D. Ward and other researchers to define an additional *Pseudokossmaticeras tercense* Zone between the *Nostoceras hyatti* (s.s.) and *Pachydiscus epiplectus* zones (Ward and Kennedy, 1993; Hancock and Kennedy, 1993; Kennedy and Hancock, 1993). They established the following succession of zones in the Tercis section (from the base upward): *Pseudokossmaticeras tercense, Pachydiscus epiplectus, Anapachydiscus fresvillensis*, and *A. terminus* (Table 3). These zones are included in the proposed global scale of the Maastrichtian Stage.

Dissimilarly to all authors who considered the Maastrichtian Stage as beginning with the *Pachydiscus neubergicus* or correlative *Pachydiscus epiplectus* Zone (both ammonite species appear in the Member N_3 of the Tercis section), we correlate the lower boundary of the Maastrichtian Stage with the base of the *Pseudokossmaticeras tercense* Zone and, correspondingly, the top of the *Nostoceras hyatti* (s.s.) Zone. With such a position, the boundary corresponds approximately to the base of the "Russian Maastrichtian" in terms of A.D. Arkhangel'skii who correlated its base with the first appearance level of belemnites from the *Belemnella lanceolata* group that was correct in opinion of Naidin (1996, p. 52). This stratigraphic level is located above the base of the *Belemnella licharewi* Acmezone (Naidin *et al.*, 1984) that was accepted for the Maastrichtian lower boundary in an early work by Jeletzky (1948) who changed his opinion later on (Jeletzky, 1958). The *Belemnella licharewi* Zone corresponds to the *Angulogavelinella gracilis* Zone of benthic foraminifers (Naidin *et al.*, 1984) that, in turn, is correlative, according to Beniamovskii and Kopaevich (1998), with the West European *Bolivinoides peterssoni–Globorotalites hiltermani* Zone located in the upper part of the terminal upper Campanian *Micraster grimmensis–Cardiaster granulosus* Zone of Germany (Schönefeld and Barnett, 1991). Based on these data, we correlate the *Belemnella licharewi* Zone with the upper part of the *Nostoceras hyatti* ammonite zone (Table 3).

It should be noted that most members of the Maastrichtian Working Group of the Subcommission on the Cretaceous Stratigraphy supported (28 pro, 8 contra, 5 abstentions), during the three-step voting at the meeting in Brussels on September 13, 1995, the GSSP fixation at the base of the *Pachydiscus neubergicus* ammonite zone. However, recognition of this level in the Boreal realm meets extreme difficulties, as is demonstrated by Christensen *et al.* (2000) who tried to trace it, using echinoids of the *Echinocorys* genus, from the Tercis section through Norfolk of England, Kronsmoor of northwestern Germany, Denmark, Poland, and Lviv depression of Ukraine. He established that *Pachydiscus neubergicus* (Hauer) occurs only in the *Belemnella sumensis* Zone of the European paleobiogeographic region (Austria, Denmark). The appearance of this species at lower levels (inside the *Belemnella lanceolata* Zone) in the Vistula River valley and Nagoryany site (Lviv depression) was only assumed, and correlation of the Tercis section with classical successions of the Boreal Province is very problematic therefore.

In the proposed scale, where the lower boundary of the Maastrichtian Stage corresponds to the base of the *Pseudocossmaticeras tercense* Zone, the *Nostoceras hyatti* and *Didymoceras donezianum* zones below this level are referred to the Campanian Stage and represent its terminal part above the *Bostrychoceras polyplocum* Zone. Following recommendations of the working group, we subdivide the Maastrichtian Stage into two substages, placing boundary between them at the base of the *Anapachydiscus fresvillensis* Zone.

PLANKTONIC FORAMINIFERS

The planktonic foraminifer zonation elaborated by Caron (1985) later was modified and specified by Robaszynski and Caron (1995). Verified in the Mediterranean Province, this zonation is appropriate for the Cenomanian–Turonian interval in the Boreal Province as well. It is very important that the foraminiferal scale is reliably correlated with magnetostratigraphic succession and controlled by isotopic age determinations of stage boundaries. Correlation of zonal units with the ammonite scale was performed by Robaszynski and then corrected using the results of recent studies (Hardenbol *et al.*, 1998).

The base of the *Rotalipora globotruncanoides* Zone, based on which the working group suggests to establish the GSSP of the Cenomanian Stage lower boundary, is established in the Mont Risou section 6 m below the first appearance level of *Mantelliceras mantelli* (Sow.) that is accepted in our scale for the Cenomanian base (Tröger and Kennedy, 1996). According to Robaszynski and Caron (1995), the lower boundary of the overlying *Rotalipora reicheli* Zone is located slightly above the base of the middle Cenomanian, inside the *Turrilites costatus* ammonite subzone (Hardenbol *et al.*, 1998). The *Rotalipora cushmani* Zone ranges, according to their conclusion, up to the second half of the upper Cenomanian *Metoicoceras geslinianum* ammonite zone of the Boreal Realm. The overlying *Whiteinella archaeocretacea* Zone is bounded by the extinction level of index species of the underlying zone at the base and by the first appearance of *Helvetoglobotruncana helvetica* (Bolli) at the top. In the Cenomanian–Turonian section near Pueblo (USA), the last event is registered inside the *Watinoceras devonense* ammonite zone (Bengtson *et al.*, 1996). The *Helvetoglobotruncana helvetica* foraminifer zone corresponds to the lower and, partly, middle Turonian and ranges up to the middle of the *Romaniceras kallesi* ammonite zone (Hardenbol *et al.*, 1998). Robaszynski and Caron (1995) reported about occurrence of rare primitive forms of index species in underlying layers beginning from the *Rotalipora cushmani* Zone. The lower boundary of the next, *Marginotruncana sigali* Zone corresponds to the top of the *Helvetoglobotruncana helvetica* and, correspondingly, to the middle of the *Romaniceras kallesi* ammonite zone. The upper boundary of the former is marked by the first appearance of *Dicarinella concavata* and corresponds to the base of the *Romaniceras deverianum* ammonite zone (Robaszynski and Caron, 1995).

The position of the upper boundary of the overlying *Dicarinella concavata* Zone is debatable. Robaszynski and Caron (1995) placed it slightly above the base of the lower Santonian like Wagreich (1992) who studied the Austrian classical Gosau section (Northern Lime-

Table 3. Zonal subdivision of the Campanian and Maastrichtian deposits of West Europe (modified after Küchler, 2000)

Northern Aquitaine (Kennedy, 1986)			Northern Spain, southern France (Biscay area) (Ward and Kennedy, 1993)		France, Tercis section, Landes (Hancock *et al.*, 1993; Hancock and Kennedy, 1993)			Northern Spain, Navarra (Küchler, 2000)	Poland, middle course of the Vistula River (Blaszkiewicz, 1980; Kennedy *et al.*, 1992)	Northern Germany, Lower Saxony (Küchler, 2000)	Northern Germany (Ernst *et al.*, 1979; Schultz, 1985)		East European platform (Olfer'ev and Alekseev, in press)
Members after G. Arnou			Maastrichtian	*Pachydiscus epiplectus*	Maastrichtian	*Pachydiscus epiplectus*	Members after Hancock *et al.*, 1993 N_3–O	*Pachydiscus neubergicus Pachydiscus epiplectus*	*Belemnella lanceolata*	?	*Belemnella lanceolata*		*Belemnella sumensis*
				Pseudokossmaticeras tercense	Upper Campanian	*Pseudokossmaticeras tercense*	M, N_1, N_2	?					*Belemnella lanceolata*
						Nostoceras hyatti	G, H, J, K, L	*Nostoceras hyatti*	*Nostoceras hyatti*		*Micraster grimmensis–Cardiaster granulosus*		*Belemnitella lichawewi/Micraster grimmensis*
													Belemnitella langei najdini/Micraster grimmensis
Upper Campanian	*Bostrychoceras polyplocum*	Q						*Didymoceras archiacianum*	*Didymoceras archiacianum*		*Belemnitella langei*		*Belemnitella langei langei/Didymoceras donezianum*
		P_3						*Trachyscaphites pulcherrimus*	*Bostrychoceras polyplocum*	*Hoploscaphites greenlandicus Jeletzkytes compressus*	*Bostrychoceras polyplocum Jeletzkytes compressus Belemnitella minor*		*Belemnitella langei minor/Bostrychoceras polyplocum*
								Bostrychoceras polyplocum		*Bostrychoceras polyplocum*			
	Hoplitoplacenticeras marroti	P_3						*Pseudoxibeloceras phaleratum*	*Pseudoxibeloceras phaleratum*	*Pseudoxibeloceras phaleratum*	*Trachyscaphites spiniger*	*Galerites roemeri*	*Hoplitoplacenticeras coesfeldiense/Belemnitella mucronata mucronata*
								Trachyscaphites spiniger		*Trachyscaphites spiniger Patagiosites stobaei*		*Patagiosites stobaei*	
												Galeola basiplana	
								Hoplitoplacenticeras marroti	?	*Hoplitoplacenticeras* spp.	*Echinocorys conica–Belemnitella senior*		

stone Alps) and reported on first specimens of *Dicarinella asymmetrica* (Sigal), which serves as an index species of the overlying zone, above the first finds of the Santonian ammonite species *Texanites quinquenodosus* (Redten.). In the southern and western Carpathians, the boundary between *Dicarinella concavata* and *Dicarinella asymmetrica* zones is correlated with the base of the middle Coniacian (Melinte, 1999). The Working Group on the Santonian Stage certified that the assemblage of *Sigalia defloensis* (Sigal), *S. carpathica* Salaj, and *Dicarinella asymmetrica* (Sigal) occurs above the base of the Santonian Stage (Kauffman *et al.*, 1996), and we selected the variant proposed by Wagreich, Robaszynski and Caron.

The extinction level of *Dicarinella asymmetrica* (Sigal) was previously correlated with the base of the Campanian Stage (Caron, 1985). Later, Wagreich (1992) demonstrated co-occurrence of its index species with the early Campanian ammonite species *Placenticeras bidorsatum* (Roem.) in the Gosau section, based on which the lower boundary of the overlying *Globotruncana elevata* Zone was placed in the middle of the lower zone of the Campanian Stage. The base of the next, *Globotruncana ventricosa* Zone is not exactly correlated with the ammonite scale so far and, following Robaszynski and Caron, we place it arbitrarily slightly above the first appearance level of *Bostrychoceras polyplocum* (Roem.) and the top above the base of the *Didymoceras donezianum* ammonite zone. The *Globotruncanita calcarata* Zone synchronous with the nannoplankton zone CC22 corresponds to the basal part of the *Nostoceras hyatti* (s.l.) Zone in the Tercis section and to the upper part of the *Didymoceras donezianum* Zone of Poland (Blaszkiewicz, 1980; Hancock *et al.*, 1993). The upper boundary of this zone coincides with the base of the *Nostoceras hyatti* ammonite zone. The upper boundary of the overlying *Globotruncanella havanensis* Zone coincides with the appearance level of index species of the higher *Globotruncana aegyptiaca* Zone recorded in the lower part of the *Pseudokossmaticeras tercense* ammonite zone (Bralower *et al.*, 1995; Burnett, 1998). The upper part of the latter encloses the lower boundary of the next, *Gansserina gansseri* Zone of planktonic foraminifers. Finally, the base of the terminal Maastrichtian *Abathomphalus mayaroensis* Zone corresponds approximately with the boundary between the lower and upper Maastrichtian substages, being located in the lower part of the latter.

CALCAREOUS NANNOPLANKTON

Two parallel scales based on calcareous nannoplankton distribution and included into the proposed global scale are oceanic (Roth, 1978) and Mediterranean (Perch-Nielsen, 1985) ones. The latter is based on the detailed zonal scale elaborated by Sissingh (1978) for Europe and Tunisia. Nowadays, the scheme by Sissingh who defined 17 zones in the Upper Cretaceous (CC10–CC26) is widely acknowledged. Nevertheless, the oceanic scheme is also of value, because Roth who grouped the Upper Cretaceous zones (NC10–NC22) into oceanic stages (which, in his opinion, are significantly more complete than traditional continental stages) correlated them with the standard magnetostratigraphic scale of the Gubbio section (Italy). Perch-Nielsen (1985) correlated both scales with each other and their zones and subzones with units of the ammonite scale. We introduced additional corrections into her correlation scheme based on the recently published works (Wagreich, 1987, 1992; Schönefeld and Burnett, 1991; Burnett *et al.*, 1992; Hancock *et al.*, 1993; Premoli Silva and Sliter, 1995; Hardenbol *et al.*, 1998; Burnett, 1998).

Although the calcareous nannoplankton zonations are widely used in practice, sphere of their application is restricted according to recent evidence. The restrictions are related to significant changes in stratigraphic ranges of the same species in different latitudinal zones. Cold-water forms usually appear in northern latitudes earlier and disappear later as compared with warm-water species, although there are opposite relationships as well.

The base of the Cenomanian Stage does not coincide with the boundary between nannofossil zones. The lower part of the Cenomanian corresponds to the upper half of the *Eiffellithus turriseiffellii* Zone (CC9), the top of which is located slightly above the lower–middle Cenomanian boundary (Hardenbol *et al.*, 1998). The remainder of the Cenomanian Stage corresponds to the *Microrhabdulus decoratus* Zone (CC10) that terminates slightly below the Turonian lower boundary. In the Rock Canyon section, it is registered in the upper part of the *Nigericeras scotti* Zone (Bengtson *et al.*, 1996; Hardenbol *et al.*, 1998).

The Turonian Stage includes two nannoplankton zones. The lower *Quadrum gartneri* Zone (CC11) spans the lower and, partially, middle Turonian substages up to the middle of the *Romaniceras ornatissimum* ammonite zone (Hardenbol *et al.*, 1998), although Burnett (1998) places its upper boundary at the base of the middle Turonian. The overlying *Lucianorhabdus maleformis* Zone (CC12) corresponds to a summary interval of the upper middle and almost entire upper Turonian, having its upper boundary located in terminal layers of the *Subprionocyclus neptuni* ammonite zone (Hardenbol *et al.*, 1998). It should be emphasized that such position of the CC12 upper boundary is not universally accepted. Par example, Sissingh (1978) and Perch-Nielsen (1985) place it slightly above the base of the Coniacian Stage, whereas Burnett (1998) draws it at the base of the middle Turonian *Romaniceras ornatissimum* ammonite zone. We preferred the first variant.

The *Marthasterites furcatus* Zone (CC13) of the Boreal Realm corresponds in our scheme mainly to the

Forresteria petrocoriensis ammonite zone or to the lower Coniacian Substage. In the Tethyan Realm, this unit is of significantly wider (lower and middle Coniacian) stratigraphic range. In Austria (Gosau), the zone corresponds to two lower ammonite zones of the Coniacian Stage (Wagreich, 1992). In the Carpathians (Melinte, 1999), Zone CC13 ranges from the terminal upper Turonian up to the lower Coniacian inclusive, like in the proposed scale.

The *Micula staurophora* Zone (CC14) of the Boreal Realm corresponds to the middle and upper Coniacian coupled with a basal part of the lower Santonian (Perch-Nielsen, 1985; Burnett, 1998).

The *Reinhardtites anthophorus* Zone (CC15) spans the lower Santonian without its terminal part. Its upper boundary coincides with the base of the *Dicarinella asymmetrica* Zone of planktonic foraminifers (Hardenbol *et al.*, 1998).

The *Lucianorhabdus cayeuxii* Zone (CC16) corresponds to the upper lower and lower upper Santonian (Wagreich, 1992; Hardenbol *et al.*, 1998).

The *Calculites obscurus* Zone (CC17) was initially referred to the Santonian (Sissingh, 1978), but Perch-Nielsen (1985) transferred it to the basal Campanian. As was later established (Wagreich, 1992), Zone CC17 comprises the Santonian–Campanian boundary layers according to Santonian *Placenticeras polyopsis* (Dujard.) and Campanian *Placenticeras bidorsatum* (Roem.) ammonites found in Austrian Alps in the lower and upper parts of the interval, respectively.

The *Aspidolithus parcus* Zone (CC18) begins in the lower Campanian *Placenticeras bidorsatum* Zone (Wagreich, 1992; Hancock and Gale, 1996) slightly above the base of the *Globotruncanita elevata* Zone of planktonic foraminifers (Wagreich, 1992). The same base level of Zone CC18 was established by K. von Salis (Hardenbol *et al.*, 1998) in the Tethyan Realm. In the Boreal Province, however, she places it lower, at the base of the upper Santonian, provided the three-unit subdivision of this stage. The exact position of the CC18 upper boundary is undetermined. Wagreich (1992) suggests it inside the lower Campanian *Globotruncanita elevata* Zone of planktonic foraminifers, whereas K. von Salis (Hardenbol *et al.*, 1998) considers it as coinciding with the lower–upper Campanian boundary in the Tethyan Realm. In our scale, we accepted the first variant.

The *Calculites ovalis* Zone (CC19) corresponds to the interval of the uppermost lower Campanian (terminal layers of the *Delawarella campaniensis* ammonite zone) coupled with the upper Campanian *Hoplitoplacenticeras marroti* and *Bostrychoceras polyplocum* (basal portion) ammonite zones. Its upper boundary is located slightly above the base of the *Globotruncana ventricosa* Zone of planktonic foraminifers (Hardenbol *et al.*, 1998).

According to Burnett (Hancock *et al.*, 1993) and K. von Salis (Hardenbol *et al.*, 1998), the *Ceratolithoides aculeus* Zone (CC20) of the Tethyan Realm corresponds to the lower part of the *Bostrychoceras polyplocum* ammonite zone and *Globotruncana ventricosa* planktonic foraminifer zone. We follow this variant of correlation, because the exact position of the CC20 upper boundary and the range of the younger Zone CC21 are established without confidence in the Boreal Realm.

The *Quadrum nitidum* Zone (CC21) ranges from the middle of the *Bostrychoceras polyplocum* ammonite zone to the lower portion of the *Didymoceras donezianum* Zone (Hardenbol *et al.*, 1998). The upper boundary of the zone almost coincides with the base of the *Globotruncanita calcarata* Zone of planktonic foraminifers.

The *Quadrum trifidum* Zone (CC22) corresponds, according to Burnett (Hancock *et al.*, 1993), to the upper part of the *Didymoceras donezianum* ammonite zone and coincides in range with the *Globotruncanita calcarata* Zone of planktonic foraminifers (Hardenbol *et al.*, 1998).

The *Tranolithus phacelosus* Zone (CC23) spans the Campanian–Maastrichtian boundary layers. It corresponds to the *Nostoceras hyatti* (Campanian) and *Pseudokossmaticeras tercense* (Maastrichtian) ammonite zones and is correlated with the *Globotruncanella havanensis, Globotruncana aegyptiaca*, and lower *Gansserina gansseri* zones of planktonic foraminifers (Caron, 1985; Bralower *et al.*, 1995; Hardenbol *et al.*, 1998). Burnett (1998) correlated arbitrarily the upper boundary of the CC23 Zone with the base of the *Pachydiscus epiplectus* ammonite zone. The zone includes subzones CC23a and CC23b, boundary between which is determined by the extinction level of *Aspidolithus parcus* (Stradn.) Noel located inside the *Pseudokossmaticeras tercense* ammonite zone, i.e., slightly below the upper boundary of the *Globotruncana aegyptiaca* Zone of planktonic foraminifers.

The *Reinhardtites levis* Zone (CC24) corresponds, according to recent data (Burnett, 1998), to a part of the *Pachydiscus epiplectus* ammonite zone. K. von Salis (Hardenbol *et al.*, 1998) assumes that such range established in the Tethyan sections can be wider in the Boreal Realm, spanning through the lower part of the *Anapachydiscus fresvillensis* ammonite zone as well. The latter corresponds to the *Arkhangelskiella cymbiformis* Zone (CC25) (Perch-Nielsen, 1985; Hardenbol *et al.*, 1998) consisting of three subzones defined by the following events: CC25a by extinction of *Reinhardtites levis* Prins et Siss., CC25b by appearance of *Litharaphidites quadratus* Braml. et Mart., and CC25c by appearance of *Micula murus* (Mart.) Bukry.

The Upper Cretaceous sequence terminates with the *Nephrolithus frequens* Zone (CC26) that corresponds to

the upper *Anapachydiscus fresvillensis* and *Anapachydiscus terminus* ammonite zones (Hardenbol *et al.*, 1998). The index species of Zone CC26 was a cold-water organism and appeared significantly lower of the indicated level in high latitudes of the southern and northern hemispheres. This situation complicates correlation.

PALEOMAGNETIC SCALE

The standard Upper Cretaceous magnetostratigraphic scale is elaborated based on the classical Gubbio section in central Italy, where the Bottaccione River canyon between Gubbio and Scheggia crosscuts a succession of Albian–Eocene hemipelagic sediments. The paleomagnetic study of this section was performed by specialists headed by W.B. Alvarez (Alvarez *et al.*, 1977). The elaborated magnetostratigraphic scale was correlated with planktonic foraminifer and calcareous nannoplankton zonations in the tropical belt (Monechi and Thierstein, 1985; Premoli Silva *et al.*, 1977). Later on, subdivision of the Gubbio section based on planktonic foraminifers was substantially modified and specified (Premoli Silva and Sliter, 1995), particularly in its Santonian–Maastrichtian interval that allowed these authors to revise the former correlation between zonation schemes of planktonic foraminifers and calcareous nannofossils, to coordinate zonal subdivisions with units of the paleomagnetic scale, and to correct their interrelationships. Premoli Silva and Sliter used the integrated nannoplankton-based scale, combining that of Sissingh for the Cenomanian–Santonian interval and that of Roth for the Campanian–Maastrichtian. In essence, the same scale is published in fundamental works *Geochronology, Time Scales, and Global Stratigraphic Correlation* (Bralower *et al.*, 1995; Gradstein *et al.*, 1995) and *Mesozoic and Cenozoic Sequence Chronostratigraphic Framework of European Basins* (Hardenbol *et al.*, 1998).

The most part of the Upper Cretaceous corresponds to the Subchronozone C34N (Aptian–Santonian) of normal polarity that terminates, in opinion of Van Hinte (1976), the Mercanton Superchron of the Late Cretaceous Quiet Magnetic Epoch of normal polarity. The polarity reversal (near the C34N–C33R boundary) coincides with the boundary between *Dicarinella asymetrica* and *Globotruncanita elevata* foraminiferal zones, which was previously correlated with the Santonian–Campanian boundary. As was mentioned, the changes in composition of planktonic foraminifers are established above the first appearance of the early Campanian ammonite species *Placenticeras bidorsatum* (Roem.). Therefore, the upper boundary of the Mercanton Superchron is located slightly above the base of the Campanian Stage. However, there is another opinion: in southeastern England, the polarity reversal occurs in layers with *Uintacrinus socialis*, i.e., in the terminal part of the upper Santonian (Hancock and Gale, 1996). Age relationships of subchrons C34N and C33R with zones of planktonic foraminifers and nannofossils are unknown so far.

In its stratigraphic range, Subchron C33R corresponds to the *Globotruncanita elevata* Zone of planktonic foraminifer, and its upper boundary coincides with the base of the *Globotruncana ventricosa* Zone (Premoli Silva and Sliter, 1995; Robaszynski and Caron, 1995). According to data from these works, the overlying normal-polarity subchron (C33N) corresponds to the interval of *Globotruncana ventricosa, Globotruncanita calcarata*, and *Globotruncanella havanensis* zones; its upper boundary is located slightly below the top of the latter zone. Khramov *et al.* (1982) included Chron C33 into the terminal part of the Jelal Superchron (Aptian–Campanian), whereas Van Hinte (1976) refers it to the basal part of the Beringov Superchron of the Campanian–Quaternary age.

Chron C32 includes two subchrons of the reversed polarity ($C32R_2$ and $C32R_1$) with two conjugated normal-polarity subchrons ($C32N_2$ and $C32N_1$). The lower part of this chron corresponds to the uppermost *Globotruncanella havanensis* and *Globotruncana aegyptiaca* zones, whereas its upper part is correlative with the lower portion of the *Gansserina gansseri* planktonic foraminifer zone. The boundary between two last foraminiferal zones falls in the upper part of Subchron $C32N_2$. Khramov with colleagues begins the Central Asian Hyperchron (Maastrichtian–Quaternary) with Chron C32.

Chron C31 is represented by the reversed-polarity (C31R) and normal-polarity (C31N) subchrons. The first of them corresponds to the upper part of the *Gansserina gansseri* Zone and the second one to the *Abathomphalus mayaroensis* Zone of planktonic foraminifer.

Chron C30 also consists of two subchrons, but its lower reversed-polarity subunit (C30R) includes a normal-polarity event that is not shown in the magnetostratigraphic scale because of a minor scale. The Maastrichtian Stage is crowned by the lower part of Subchron C29R. Similarly to Chron C30, this part of the section corresponds to the *Abathomphalus mayaroensis* Zone of planktonic foraminifers.

Recently, new publications demonstrated existence of reversed-polarity intervals within the Cretaceous normal-polarity superchron (Tarduno *et al.*, 1992; Baraboshkin *et al.*, 1997). At the moment, their inclusion into the paleomagnetic part of our scale seems premature.

GEOCHRONOLOGICAL SCALE

The geochronological scale is taken from recent works by Gradstein *et al.*, (1995, 1999) and Hardenbol *et al.*, (1998). It is calibrated using the dates obtained by the $^{40}Ar/^{39}Ar$ method for sanidine from bentonite interbeds in the Upper Cretaceous sections

of the Western Interiors, United States. These dates are extrapolated to the European sections using the ammonite-based detailed correlation proposed in a series of works (Cobban, 1993; Obradovich, 1993; Hicks, 1993; Gradstein *et al.*, 1999). The only modification concerns the Campanian–Maastrichtian boundary, because both stages have different ranges in North America and Russian. In accordance with recommendation of the First International Symposium (Copenhagen) to place the Campanian–Maastrichtian boundary at the base of the *Belemnella lanceolata* Zone, it is correlated with the boundary between the *Globotruncanita calcarata* and *Globotruncanella havanensis* zones in the Gubbio section (Bralower *et al.*, 1995). Based on paleomagnetic data, this boundary was recognized in the Western Interiors of United States (Cande and Kent, 1992), where its radiological age is estimated to be 74.5 Ma (Gradstein *et al.*, 1995). Inasmuch as the Campanian–Maastrichtian boundary is placed inside the *Globotruncanella havanensis* Zone of the proposed global scale, duration of which is 1.2 m.y. (Hardenbol *et al.*, 1998), we consider arbitrarily the base of the Maastrichtian Stage as corresponding in age to 74.0 Ma. In the scale accepted by Gradstein *et al.* (1995, 1999), this boundary is placed at the base of the North American *Baculites eliasi* Zone; this level is correlated with the lower boundary of the European *Pachydiscus neubergicus* Zone and dated at 71.3 Ma. Age values accepted for the Upper Cretaceous stage boundaries in the proposed scale slightly differ from those recommended by the Interdepartmental Stratigraphic Committee of Russia (Rublev, 2000).

CONCLUSION

The scale characterized above represents, in our opinion, an appropriate tool for direct or indirect correlation of regional scales in all main regions of Russia, where Upper cretaceous deposits are widespread, e.g., in the Russian platform, North Caucasus, West Siberia, and the Far East. Nevertheless, the scale can be specified after appearance of new data.

First, some modifications are unavoidable at the final approval of stratotypes of stage boundaries by the International working groups. The Cretaceous System needs, unlike some others, a universal understanding of stage boundaries. After their approval by the International Union of Geological Sciences, the general scale of Russia should be properly corrected.

Second, further investigation of calcareous nannofossils may reveal some inaccuracies in correlation of zonal successions in the accepted scale. We may expect as well new data on the Upper Cretaceous ammonites of the Russian platform after the purposeful sampling that was interrupted long ago.

Despite all possible disadvantages, the global scale of the Upper Cretaceous is a necessary tool for practical stratigraphy, long-expected by Russian geologists.

Reviewer M.A. Akhmet'ev

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Upper Cretaceous Deposits in Eastern Piedmonts of the Lesser Caucasus

T. Ab. Gasanov*†, O. G. Melikov, Sh. A. Babaev*, and G. A. Aliev***

*Institute of Geology, Azerbaijan Academy of Sciences, Baku, Azerbaijan

**Azerbaijan State Academy of Petroleum, Baku, Azerbaijan

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Abstract—The Upper Cretaceous sequence in the east of the Lesser Caucasus, within the Azerbaijan territory, begins with the Cenomanian terrigenous sediments that are overlain with a hiatus by thick volcanogenic–terrigenous deposits of the Coniacian and Santonian stages. The Campanian–Maastrichtian interval and lower Paleocene are composed mainly of limestones bearing echinoids, inocerams, and planktonic foraminifers, which are used to distinguish the substages. Boreholes between the Agdam and Fizuli cities and near the town of Terter recovered the Maastrichtian and lower Paleocene limestones below the Akchagylian cover.

Key words: Upper Cretaceous, stratigraphy, echinoids, inocerams, ammonites, foraminifers, orbitoids, Lesser Caucasus, piedmonts.

INTRODUCTION

The Upper Cretaceous deposits in eastern piedmonts of the Lesser Caucasus are rather widespread and represented by volcaniclastic, pyroclastic, and carbonate sediments.

Being easily accessible, exposures of the Upper Cretaceous deposits in this region are studied much better than elsewhere in the Lesser Caucasus (Figs. 1, 2). In various periods, they have been investigated by Khalilov (1959), Melikov (1966), Shikhalibeili (1966, 1996), Aliev (1967), Mamezade (1967), Khalilov *et al.* (1974), Aliyulla (1977), Alizade *et al.* (1988), and Gasanov (1985, 1988, 1996).

Geological survey and special stratigraphic investigations that were performed in 1990–1996 yielded many new data on distribution and stratigraphy of the regional Upper Cretaceous, which are considered here with due regard for results of earlier works.

The distribution area of Upper Cretaceous deposits in eastern piedmonts of the Lesser Caucasus extends for 300 km and includes the Shikhly–Kazakh–Kovlyar, Dzegam–Khanlar–Giyasly, and Agdam–Fizuli–Dzhebrail zones.

GEOLOGY AND TECTONICS OF THE LESSER CAUCASUS PIEDMONTS

In terms of tectonics, the distribution area of Upper Cretaceous deposits is confined to subsiding parts of the Gazakh–Agdzhakend and Agdara[1] depressions in the Lok–Agdam zone[2] and of the Khodzhavend[1] depression in the Gekcha–Karabakh zone of the Lesser Caucasus (Fig. 2).

Materials presented below are obtained from 43 geological sections, in which we collected macro- and microfauna (echinoids, mollusks, and foraminifers) during the last decade, and characterize the lithology and paleontology of 54 boreholes drilled in the eastern piedmonts of the Lesser Caucasus. Below thick molasse deposits, these boreholes recovered the Upper Cretaceous rock rich in macro- and microfossils (Figs. 3, 4).

Deposits of the Cenomanian Stage (Damirla Formation 250 m thick) are most widespread in the Dzhogas–Indzhasu interfluve and at the Mt. Karvakar site (areas I and II). They correspond to rhythmical alternation of sandstones, siltstones, tuffaceous shales, and conglomerates (150–250 m). The assemblage of Cenomanian foraminifers includes *Rotalipora turonica turonica* (Brotz.), *Praeglobotruncana stephani agdavanica* Alij., *Talmanninella greenhornensis* (Brtz.) (determined by S.I. Suleimanova and R.M. Alieva). Macrofossils are represented by *Amphidonte columba plicatula* Lam., *A. columba chaperi* Bayle, *A. columba silicea* Lam., *Nucula bagirliensis* O. Aliev (in coil.), *Variamussium ninae* (Karak.), *Oligoptyxis* ex gr. *turricula* Pčel., *Itruvia caucasica* Pčel., *Avellana dubia* Briart et Cornet, *Acanthoceras rhotomagense* Schloth., and *Neohibolites ultimus* d'Orb. (determined by

[1] Formerly known as Mardakert synclinorium.

†Deceased.

[2] Gasanov (1986, 1996) who was adherent of tectonic scheme by L.N. Leont'ev and V.E. Khain contradicting that of Shikhalibeili (1996) distinguished the Lok–Karabakh zone instead of Lok–Agdam zone, and the Gekcha–Akera zone instead of Gekcha–Karabakh zone.

O.B. Aliev). The listed forms have been collected from shales, sandstones, and clayey limestones exposed near the villages of Agdzhakend, Aganus, Chorman, Kushchi-Airum, Baganis, Koshkotan, Kotigekh, Dostlu, and Mazamli River southward of the town of Shamkir.

The reliably dated Turonian deposits are unknown in the eastern part of the Lesser Caucasus. Cenomanian sediments conformably overlie here the upper Albian sandy shales. In turn, they are discordantly overlain by basal conglomerates of the Coniacian tuffite sequence.

Coniacian Stage. Voncanogenic–sedimentary and carbonate–terrigenous deposits of this stage are widespread in the Gazakh–Agdzhakend synclinorium, where their sequence is 670–800 m thick. Lithological features and fossils allowed us to distinguish here the lower Coniacian and uper Coniacian–lower Santonian subdivisions.

Volcanogenic–sedimentary deposits of the lower Coniacian Substage (670–800 m) are traceable within a narrow belt extending from the Tauzchai River in the northwest to the right-hand side of the Bala-Kyurakchai River in the southeast (areas I and II). Along this profile, basaltic andesites grade into andesites and dacites. In the Tauzchai–Akstafachai interfluve, the dominant rocks are tuffaceous conglomerates and breccias, whereas in the east (Dzegamchai–Shamkirchai interfluve, area II) andesitic lavas represent occasional components of the sequence, and pyroclastic rocks are intercalated with carbonate members (40–60 m thick). Toward the southeast, abundance of pyroclstic interlayers gradually increases again, and andesitic to dacitic lavas and their subvolcanic equivalents simultaneously become occurring frequently.

In the study region, the lower Coniacian deposits transgressively overlie various rocks of the Middle–Late Jurassic, Early Cretaceous, and Cenomanian ages. The early Coniacian age of rocks in question that was originally inferred by Abdullaeva *et al.* (1963) is confirmed by results of our investigations. In addition to formerly known sites of fossil fauna, we discovered new ones near the village of Alibeili, at the left side of the Dzhogas River, and southward of the town of Shamkir. Clayey limestones and calcareous sandstones yielded here the molluscan assemblage of *Actaeonella crassa* Duj., *Radiolites galloprovincialis* Math., *Trigonoarca ligeriensis* d'Orb., *Exogyra tausensis* Renng., *Isocardia karabakhensis* (Bobk.), *Plesioptygmatus bicincta* (Bronn.), *Pl. scalaris* Pčel., and *Pl. tausensis* Pčel. According to conclusion of O.B. Aliev, the rocks bearing this assemblage are of the early Coniacian age. They can be divided into the lower volcanogenic–sedimentary (500 m) and upper pyroclastic–sedimentary (250–300 m) subformations.

The upper Coniacian–lower Santonian deposits, which are up to 1000 m thick, are distinguished in the Indzhasu–Bala-Kyurakchai interfluve (area II). Their age is defined according to the intermediate position between rock units bearing the late Coniacian and early

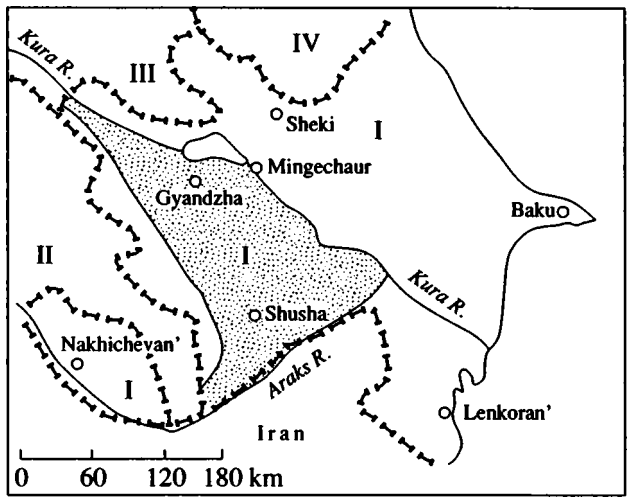

Fig. 1. Geographic position of the study area in the Lesser Caucasus: (I) Azerbaijan; (II) Armenia; (III) Georgia; (IV) Dagestan.

Santonian fossils. In addition, they yield organic remains characteristic of the Coniacian–Santonian interval (Shikhalibeili, 1964). The main constituents of deposits under consideration are terrigenous and pyroclastic rocks of andesite–dacite composition. Pyroclastic beds (1000 m) are most widespread in the Gyandzhachai River basin, where they are discordantly overlain by the upper Santonian limestones (Figs. 5, 6).

The **Santonian Stage** is divided in two substages. The lower one is composed of tuffaceous sandstones and shales, tuffites, clays, and limestones (350 m), which are exposed near the villages of Shikhly, Kamarli, Farzali, and Dagkesaman. In the Indzhasu and Gasansu river basins, these sediments transgressively overlie the Coniacian pyroclastic and sedimentary beds (Fig. 5). The early Santonian age of deposits under discussion is indicated by occurrence of foraminifers typical of the *Globotruncana concavata* Zone. These are *Globotruncana subbotinae* Alij., *G. concavata* Brotz., *G. tricarinata tricarinata* (Guer.), *G. bulloides bulloides* Vogl., *G. lapparenti* Brot., *G. linneiana* (d'Orb.), *Globigerinelloides aspere* (Ehr.), *Heterohelix santonica* (Agal.), *H. elegans* (Rzehak), *H. plummerae* (Loett.), *H. globulosa* (Ehr.), *Eponides concinnus* Bronn., *Gavelinella costulata* (Marie), *Stensioeina exculpta* (Reuss.), *S. emscherica emscherica* Barysch., *Gyroidina turgida obliquaseptata* Mjat., and *Striatella santonica* (Agal.), all determined by R.M. Alieva and discovered from clay and siltstone beds.

The upper Santonian Substage corresponds to the sequence of clayey, sandy, and gravely limestones, which are 150 m thick in total. In the Indzhasu–Akstafa interfluve, these limestones were investigated earlier by V.P. Rengarten, R.N. Abdullaev, and R.N. Mamezade and others. Here and at the Elensutapa locality, we collected mollusks *Sauvagesia meneghini* Pir., *Exogyra*

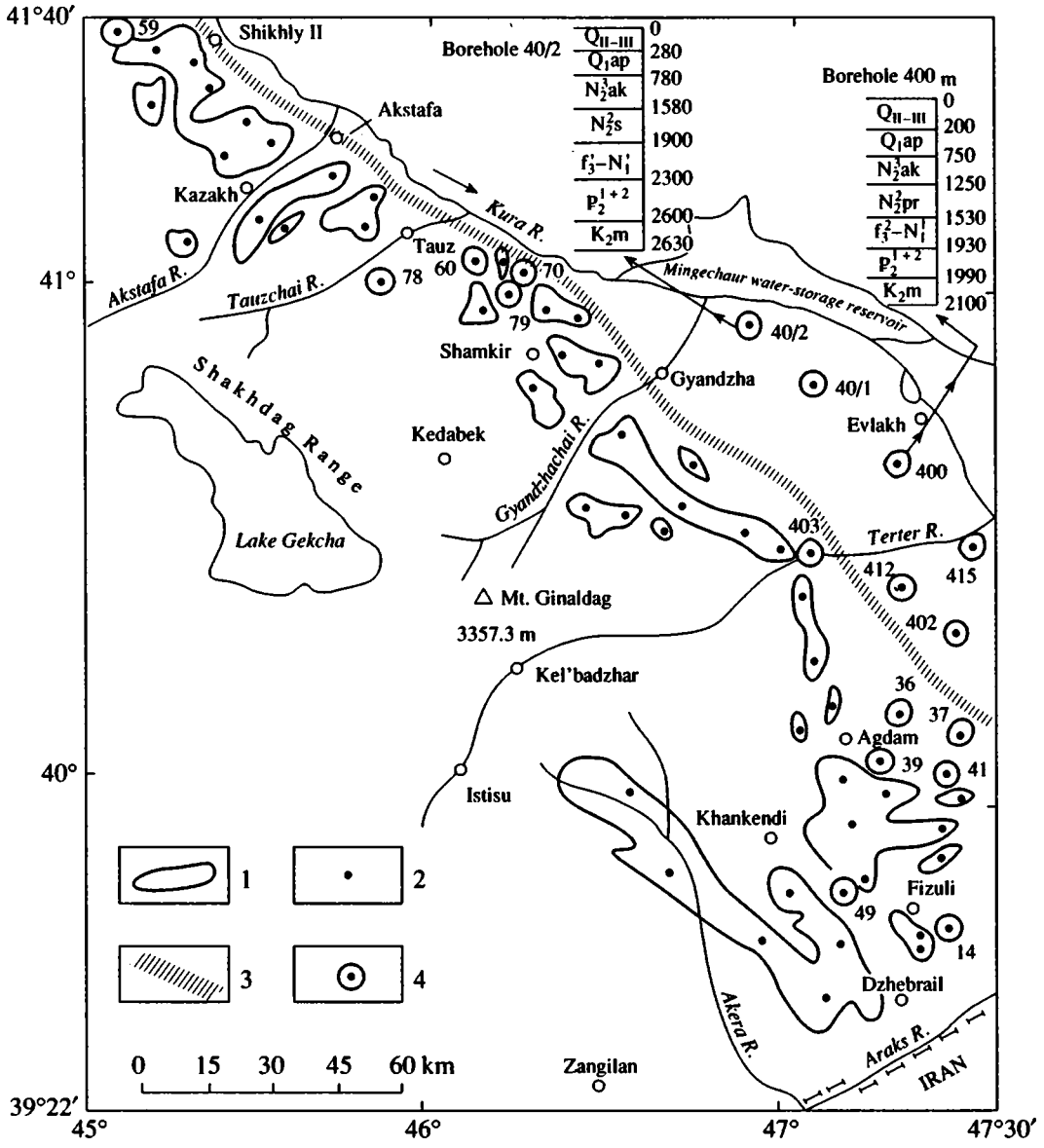

Fig. 2. Distribution areas of Upper Cretaceous deposits in eastern piedmonts of the Lesser Caucasus (sketch compiled by T.Ab. Gasanov and O.G. Melikov): (1) Upper Cretaceous limestones; (2) sites of lithological and paleontological investigations; (3) deep-seated fault detected by gravimetric data; (4) boreholes, which penetrated the Upper Cretaceous–Apsheronian deposits.

lateralis Mills., *Marsupites testudinarius* Schloth., *Conulus oblongus* d'Orb. (determined by O.G. Melikov), and foraminifers *Globotruncana subarca* Alij., *G. subventricosa* Alij., *G. subbotinae* Alij., and *Heterohelix plana* (Cushman) (determined by R.M. Alieva). In the northeastern piedmonts of the Lesser Caucasus, the upper Santonian deposits are exposed in hilly ranges of northwestern strike, which are distributed en echelon and become lower toward the Kura depression. These deposits grade upward into the Campanian limestones (Figs. 3–5).

The **Campanian Stage** (Mirushen Formation) corresponds to a sequence of carbonate facies distributed locally. Campanian sediments occur in depressions of the Lok–Agdam zone and are divided into the lower (70–120 m) and upper (150–200 m) substages. Both substages are interconnected via transitional layers everywhere. The early Campanian age of the lower unit is indicated by echinoids *Paronaster cupuliformis* Airaghi, *Micraster schroederi* Stoll., *Galeola senonensis* d'Orb., *Pseudoffaster schmidti* Moskvin, *Seunaster gillieroni* Lor. (determined by O.G. Melikov), and by planktonic foraminifers *Globotrucana arca* (Cushm.), *G. califormis* (Lapp.), *G. rosetta* (Cars.), *Heterohelix striata* (Ehr.) (determined by R.M. Alieva). Echinoids have been collected from chalklike sandy limestones and marls at the Mt. Babakar, Agdag, Aidar, and Elensutapa sites, in the Akstafa–Tauzchai interfluve, and

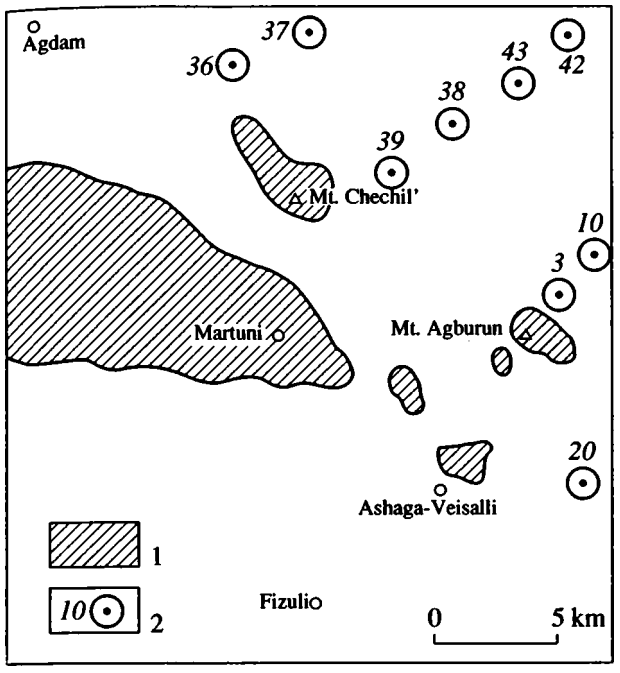

Fig. 3. Distribution areas of Upper Cretaceous deposits in the southeastern buried part of the Khodzhavend synclinorium: (1) Upper Cretaceous deposits; (2) boreholes by numbers.

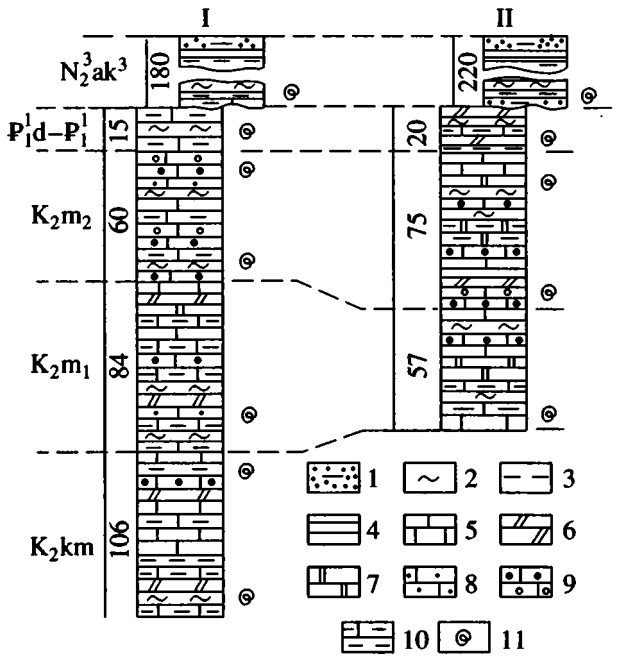

Fig. 4. Correlation scheme of Upper Cretaceous–Paleocene deposits in the southeastern buried part of the Khodzhavend synclinorium: (1) clayey sandstone; (2) shale or siltstone; (3) marl; (4) claystone; (5) medium-bedded, (6) chalklike, (7) sugary, (8) sandy, (9) gravely, and (10) clayey limestones; (11) sampling levels of fossils; (I) Mt. Chechil' and (II) Mt. Agburun sections. Indices of Arazbar ($N_2^2ak^3$), Sardarashen ($P_1^1d-P_1^1$), Dashbashi (K_2m_2), Karkarchai (K_2m_1), and Mirushen (K_2km) formations.

near the town of Shamkir. In most sections, the Campanian deposits are transgressively overlapped by Maastrichtian gravelstones and sandy sandstones.

In southeastern areas of the Lesser Caucasus, the undivided Campanian limestones are exposed at the Mt. Chechil' and Mt. Agbulak sites (earlier observations by M.D. Gavrilov, V.K. Khalilzade, A.A. Bairamov, and M.M. Zeinalov). About 0.7 km southwestward of the Mt. Chechil', exposures of these limestones are traceable southeastward up to the Agbulak Spring along the highway Agdam–Fizuli, where their beds dip northeastward (30°–60°) under angles of 10°–30°. Near the spring, where dip angles are steeper (60°–85°), the limestone beds are disturbed by reversed faults of northwestern (300°–320°) and northeastern (20°–30°) strikes. Rock samples, which have been collected by Gasanov near the spring from shale interlayers in marly limestones, yielded planktonic foraminifers *Globotruncana fornicata* Plumm., *G. arca* (Cushm.), *G. subarca* Alij., *G. ventricosa* Alij., *G. linneiana* (d'Orb.), *Heterohelix porosa* Alij., *H. globulosa* (Ehr.), *Stensioeina mursataiensis* Vass., *Marssonella* aff. *indentata* (Cushm.), and *Eponides sparksi* White. According to conclusion of R.M. Alieva, these fossils indicate the late Santonian–early Campanian age of their host deposits. Shale and marl beds located 10 m above yield the Campanian inocerams *Inoceramus balticus* Boehm, *I. azerbaidjanensis* M. Aliev, and *I. sarumensis* Woods in association with foraminifers *Globotruncana arca* (Cushm.), *G. subarca* Alij., *G. linneiana* (d'Orb.), *G. ventricosa* White, *Heterohelix globulosa* (Ehr.), *H. striata* (Ehr.), *Gyroidina turgida* Hag. (determined by R.M. Alieva).

About 50–60 m higher, close to the limestone quarry, Khalilzade (1957) found a similar assemblage of Campanian fauna. The same limestones are exposed in the southeastern termination of the Khodzhavend synclinorium in the Gekcha–Karabakh zone, where their beds plunge under recent deposits of the Lesser Caucasus foredeep.

Age of limestones at the Mt. Chechil' site was originally established by Khalilzade (1957), who discovered *Inoceramus regularis* d'Orb., *I. balticus* Boehm, and *Endocostea inflexa* Blein. of the Campanian age (determined by R.N. Mamezade) from beds situated 70–80 m above the Agbulak Spring. In addition, we collected here echinoid remains *Galeola senonensis* d'Orb., *Pseudoffaster schmidti* Moskvin, *Seunaster gillieroni* Lor., and *Micraster schroederi* Stoll. (determined by O.G. Melikov). Shikhalibeili (1964) and Zeinalov (1969) also suggested the Campanian age for these limestones, but in geological maps of the Caucasus and Azerbaijan (scale 1 : 500000) they are mistakenly attributes to the Hauterivian.

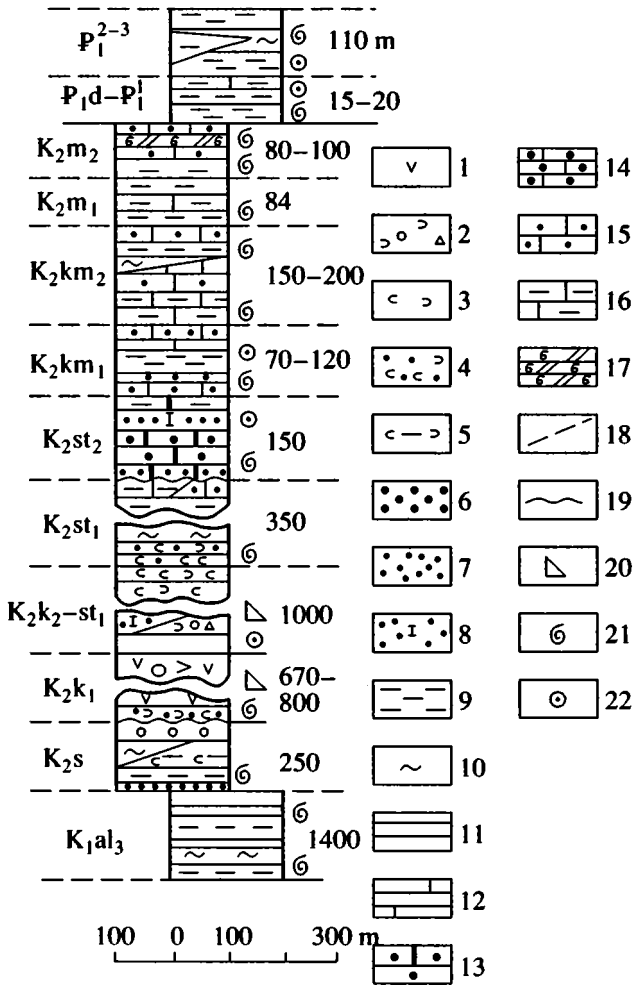

Fig. 5. Composite litho-stratigraphic column of Upper Cretaceous deposits in eastern piedmonts of the Lesser Caucasus (compiled by T.Ab. Gasanov and O.G. Melikov): (1) andesite or andesitic dacite; (2) tuffaceous conglomerate or tuff-breccia with andesite–dacite clasts; (3) tuffite; (4) volcanic gravel; (5) silty tuffite; (6) gravelstone; (7) sandstone; (8) calcareous sandstone; (9) siltstone; (10) shale; (11) clay and shale; (12) medium-bedded, (13) highly gravelly reddish, (14) gravelly, (15) sandy, (16) clayey, and (17) orbitolinid limestones; (18) conformable and (19) discordant boundaries; (20) rocks dated by K–Ar method, Ma; (21) sampling level of macrofauna; (22) sampling level of microfauna.

In the geological map of the eastern Lesser Caucasus that was compiled by Gasanov in 1983–1988 and revised in 1990 the limestones, which are exposed at the Mt. Agburun, Mt. Chechil', Ashaga-Veisalli, and Fizuli localities, are shown for the first time subdivided into the upper Santonian–lower Campanian, upper Campanian, lower and upper Maastrichtian, and lower Paleocene beds (Figs. 3, 4).

Data considered above suggest undoubtedly the Campanian age of lowermost limestone beds exposed in the Mt. Chechil' area. Characteristic rocks of the Campanian sequence are fine- to medium-grained laminated limestones and marls intercalated with thin interlayers (1–20 cm) of shales and siltstones in the lower part, and with chalklike limestone interbeds in the upper one. The sequence is correlated, based on its lithological peculiarities, with the Mirushen Formation distinguished by Gasanov and Kyazymov (1988) near the eponymous village of the Khodzhavend district. In southeastern areas of the Lesser Caucasus, the Campanian chalklike limestones grade into clayey limestones of the lower Maastrichtian (Mt. Chechil' and Mt. Ilanlydag sites, Fig. 3).

Maastrichtian Stage. Carbonate deposits of this stage are locally preserved from erosion in frontal elevations of the Lesser Caucasus piedmonts (Mt. Kyakil', Mt. Malachal, Dzhogas River basin, and others). Rengarten (1959) was first to establish their age, and his inference was later supported by Abdullaev (1963). Mamezade (1967) subdivided the Maastrichtian deposits in the Gazakh–Agdzhakend depression into lower and upper substages. Later on, investigating exposures at the Khanlar, Mikhailovka, Babakar, and Malachal localities, we found echinoid remains *Conulus magnificus* d'Orb., *Catapygus laevis* Desor, and *Coraster* cf. *vilanovae* Cott. (determined by O.G. Melikov) in association with inocerams *Inoceramus lingua* Goldf. and *I.* cf. *tegulatus* Hag. (determined by O.B. Aliev). The Maastrichtian deposits are intermittently exposed within a belt extending from the village of Shikhly in the northwest, and across the Shamkir, Dzegam, and Khanlar localities, to the area between the towns of Agdam and Fizuli in the southeast (Fig. 2). In the last area, they have been recovered by drilling.

In the eastern part of the Khodzhavend synclinorium (northeastern slopes of the Mt. Chechil' and Mt. Agburun, Fig. 4), the Campanian–Maastrichtian deposits plunge under Pleistocene deposits of the Lesser Caucasus foredeep. In addition to the indicated mountains, there are three smaller outcrops (20 to 150 m^2 in size) of these rocks, which are situated between the village of Ashaga-Veisalli and Mt. Agburun. Formerly, rocks of these outcrops were attributed to the Berriasian–Valanginian interval (Alizade *et al.*, 1973).

Two substages are distinguishable in the Maastrichtian limestone sequence of the eastern Lesser Caucasus (Figs. 3, 4). The lower one (Karkarchai Formation 102 m thick) includes beds of sandy limestones and marls, which are intercalated with thin shale interbeds. The latter yield characteristic echinoid remains *Echincorys pyramidata* Portl., *Cardiaster granulosus* Goldf., *Seunaster altus* Seunes, *Galeaster sumbaricus* Posl., *Homoeaster tunetanus* Pomel, *Guettaria rocardi* Cott., and *Cyclaster integer* Seunes (determined by O.G. Melikov, consultation by M.M. Moskvin).

The total thickness of sugarry, clayey, sandy, and gravely limestone beds and shale interlayers, which constitute the lower Maastrichtian Karkarchai Formation in all outcrops, varies from 57 (Mt. Agburun site)

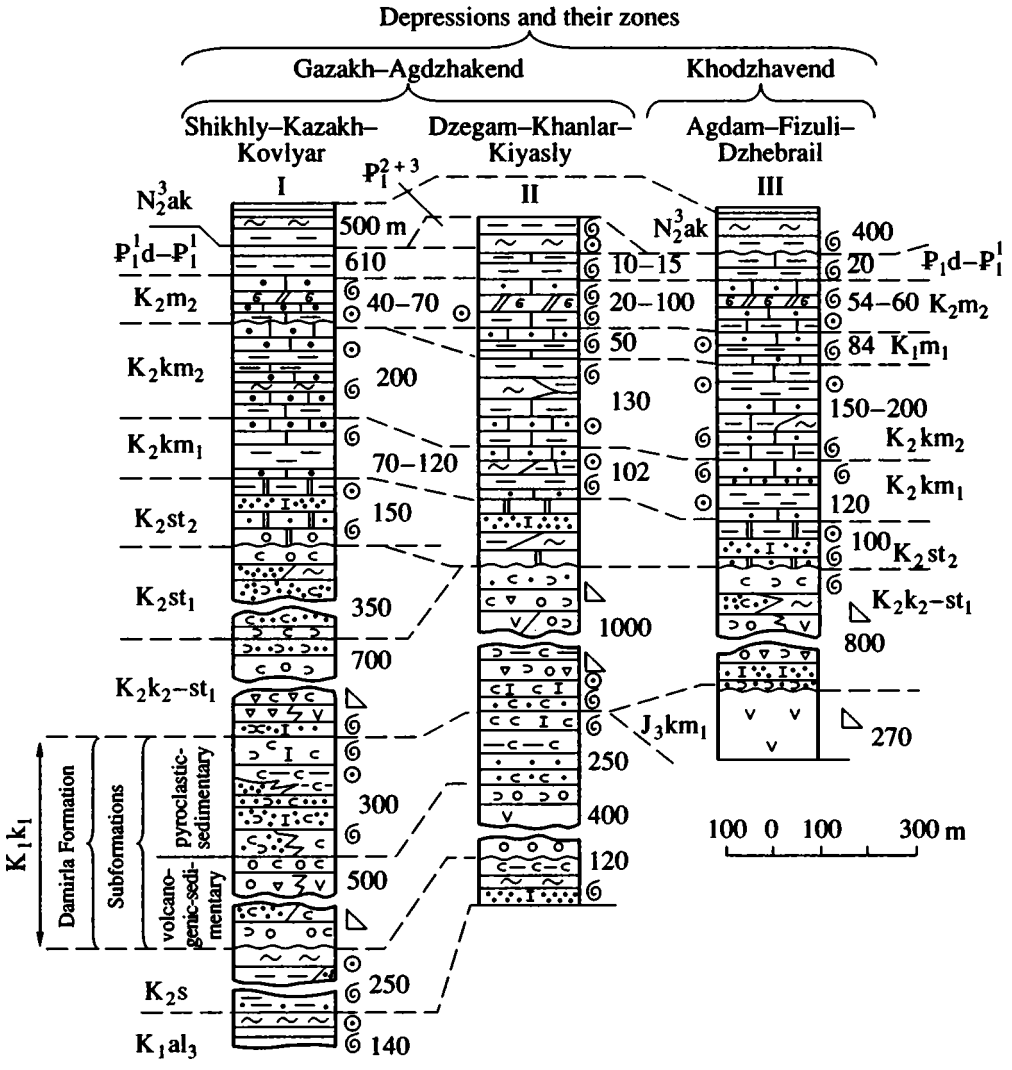

Fig. 6. Correlated sections of Upper Cretaceous deposits in eastern piedmonts of the Lesser Caucasus (symbols as in Fig. 5).

to 102 m (Khanlar site). Clayey limestones with shale intercalations, which are charactiristic of the unit, are concentrated in the lower (Mt. Chechil' and Mt. Ilanly-dag) and, sometimes, in the middle parts of the unit. Their beds gently dip (angles of 5°–10°) northeastward everywhere.

The early Maastrichtian age of deposits inder consideration is indicated by the index echinoid species *Seunaster altus* Seunes, *Homoeaster tunetanus* Pomel, *Galeaster sumbaricus* Posl., and inocerams *Inoceramus regularis* d'Orb. and *I. tegulatus* d'Orb., which have been collected from lower and middle beds of clayey to sandy limestones at the Agburun, Chechil', and Ashaga-Veisalli localities. In addition, sandy limestones exposed 1 km south-southwest of the town of Fizuli yield the early Maastrichtian brachiopods *Neoliothyrina* cf. *obesa* (Dav.), *Carneithyris carnea* Sow. (determined by M.V. Titova), and oyster *Pycnodonte vesiculare clavate* Nills. of the Campanian–Maastrichtian age (determined by V.M. Kharitonov).

Everywhere, the lower Maastrichtian limestones grade upward into the upper Maastrichtian gravely limestones of the Dashbashi Formation (54–60 m). In addition to rocks of the last type, this formation also includes middle-bedded clastic chalklike and sandy limestones intercalated with thin shale interlayers (1–5 cm).

The late Maastrichtian age of the formation is established based on echinoids *Conulus magnificus* d'Orb., *Catapygus laevis* Desor, *Echinocorys cypliensis* Lamb., *E. perconica* Hag., *E. arnoudi* Seunes, *Hemipneustes striatoradiatus* Leske, *Seunaster lamberti* Charlet, *Pseudoffaster renngarteni* Schmidt, *Homoeaster evaristei* Cott., *Coraster vilanovae* Cott., *Cyclaster integer* Seunes, which associate with solitary corals *Balanophyllia schlosseri* Traub. (determined by O.G. Melikov).

The above fossils have been collected from lower beds of the unit (Mts. Chechil' and Agburun) and near its top (outcrops eastward of the Mts. Agburun and Ilanlydag). About 1 km to the north and northeast of the Mt. Chechil', the upper part of the unit includes a member (14 m) of white and yellowish, gravely clastic limestones. This member bears in abundance the late Maastrichtian orbitoids *Orbitoides media* (D'Arch), *Z. tissoti* (Schlum.), *O. apiculata apiculata* Schlum., and *O. apiculata aplanata* Renng. (determined by Sh.A. Babaev). Beds of the member dipping northeastward under angles of 5°–10° are traceable from the southeast to the northwest for a distance of 1.5 km.

Under recent deposits, the upper Maastrichtian limestones have been recovered from depth of 150 to 210 m by boreholes drilled in northern and northeastern outskirt sectors of Agdam (boreholes nos. 36, 37, 39, 3, 10, 20) and in the northeastern suburb of Fizuli (nos. 41, 30–32, and others, Fig. 1). The limestone beds plunge toward the deep-seated fault of the Lesser Caucasus foredeep. They have been recovered from the depth of 300 m (Borehole 38) in lower courses of the Terter River, from the depth of 2490 m near the synonymous town (Borehole 403), and from the depth of 1710 m in lower courses of the Tauzchai River (Borehole 60). According to other drilling data (boreholes 412, 415, 400, 40/1, 40/2, and others), these rocks occur at the depth of 2400–3200 m in the north-northeastern part of the Lesser Caucasus foredeep (area of the Mingechaur water-storage reservoir). Accordingly we suggest that eastern piedmonts of the Lesser Caucasus, between the foredeep and right-hand side of the Kura River, have a stepwise blocky structure (Figs. 4–6). It is also important that orbitoid limestones associate everywhere in the Lesser Caucasus with light, greenish gray, gravely to sandy limestones, which occur about 12 to 24 m below the base of the Danian Stage (Figs. 1, 2).

As for stratigraphic distribution of orbitoids, Mamedov and Babaev analyzed this issue before 1974, and later on Babaev investigated their assemblages collected in 1983–1993. According to the results obtained, these researchers distinguished the local *Simplorbites gensasicus* (Leym.)–*Lepidorbitoides socialis* (Leym.) Zone that is confined in the eastern Lesser Caucasus to the upper Maastrichtian organogenic-detrital limestones (Mamedov and Babaev, 1975).

The late Maastrichtian age of limestones is confirmed by assemblage of foraminifers *Abathomphalus mayaroensis* (Bolli), *Globotruncana arca* (Cushm.), *G. postventricosa* Vass., *Heterohelix striata* (Ehr.), *Stensioeina caucasica* (Subb.), *Psedotextularia varians* (Rzeh.), *Parrella navarroana* (Cushm.), and *Cibicides bembix* (Marss.), which have been determined by R.M. Alieva in shale interlayers (1–5 cm) intercalated with these rocks. The assemblage characterizes an upper part of the sequence between clayey and sandy limestones exposed at southeastern slopes of Mts. Agburun and Ilanlydag.

The **Danian Stage** of the Lower Paleocene (Sardashen Formation) is represented by a member (15–20 m) of bluish and whitish, moderately laminated, clayey to sandy limestones intercalated with shale and marl interlayers. These beds dip under small angles (2°–5°) toward the northeast. Shale beds, which are exposed in north-northeastern slopes of Mts. Agburun and Chechil', yield the early Paleocene foraminifers *Subbotina triloculinoides* Plumm., *S. trivialis* Subb., *Globanomalina ehrenbergi* (Bolli), and *Parasubbotina varianta* (Subb.). Based on lithology and paleontological characteristics, Gasanov and Kyazymov (1988) correlated the lower Paleocene of these areas with the Sardashen Formation.

The lower Paleocene calcareous and terrigenous beds, which gently dip (angles of 5°–10°) toward the Lesser Caucasus foredeep, are transgressively overlapped by the upper Akchagylian bluish clay and sands of the Arazbar Formation. Boreholes recovered them below thick molasse deposits of the regional Apsheronian Stage, and their distribution area extends northeastward to the town of Agdzhabedy, for a distance of 20 km (Figs. 4, 5).

These transgressive interrelations between the lower Paleocene and upper Akchagylian deposits imply that the southeastern termination of the Khodzhavend synclinorium was rising up as a block during a rather long period. The Maikop deposits widespread in the northwest of the synclinorium (Khramort site) and in the southeast, where the Agdam anticlinorium is plunging under Giyasly site), allowed Gasanov an Kyazymov (1988) to suggest that, in the basement of the Lesser Caucasus foredeep, there are stepwise subsided and risen blocks of northwestern strike.

Now we would like to characterize below a section of the lower Maastrichtian–lower Paleocene deposits, which have been studied in detail 0.5 km southward of the Mt. Agburun (Fig. 5), where beds dip northeastward (60°–70°) under angles of 30°–40°.

Lower Maastrichtian (Karkarchai Formation). (1) Limestones, cherty to a minor extent, moderately bedded, with thin (1–5 cm) shale interlayers; along the strike, rocks grade into bluish clayey limestones. The apparent thickness is 8.5 m.

(2) Moderately bedded sugary limestones (0.3–0.4 m) rhythmically alternating with gravely and brecciated limestones (0.2–0.3 m); sporadic shale interlayers between limestone beds yield the early Maastrichtian assemblage of foraminifers *Globotruncana contusa* (Cushm.), *G. conica* White, *G. lappareniti* Brotz., *G. charchaputensis* Alij., *Globotruncanita stuarti* (Lapp.), *Heterohelix striata* (Ehr.), *Stensioeina praecaucasica* Vass., and *Eponides moskvini* (Kell.). The unit is 16.5 m thick.

(3) Alternating white, somewhat sandy (0.2–0.3 m) and gravely (0.1–0.2 m) limestones; intercalated shale interlayers bear the early Maastrichtian foraminifers *Globotruncana contusa* (Cushm.), *G. morozovae* Vass.,

Distribution ranges of orbitoids in the East Lesser Caucasus

Species \ Samepling sits	s. Gyuigyui	s. Subsarkis	r. Avei	s. Nerkin Oratag	s. Dashsalakhly	s. Damirchilyar	s. Tsakkar	g. Shamkir	pos. Seifali	g. Khanlar	Gora Cheil'	Gora Agburun
Orbitella apiculata depressa Renng.	+	+	–	–	–	–	+	+	+	+	+	+
Orb. Aplanata segmentoidea Renng.	–	–	–	–	–	–	–	–	–	+	+	+
Orb. apiculata aplanata Renng.	–	–	–	–	+	+	–	–	–	–	+	+
Simplorbites gensasicus (Leym.)	+	+	+	+	+	+	–	–	+	+	+	+
Lepidorbitoides minor (Schlumb.)	+	+	+	+	+	+	–	–	+	+	+	+
Lep. socialis socialis (Ieym.)	+	+	–	+	+	+	+	+	+	–	+	+
Lep. socialis regularis M. Douv.	–	–	–	+	–	–	+	+	+	–	+	+
Orbitoides media (D'Arch.)	+	+	–	+	+	+	–	–	–	–	–	–
Orb. tissoti Schlumb.	–	–	–	–	–	–	–	+	–	–	+	+
Orb. Sp.	–	–	–	+	–	–	–	–	–	–	–	–
Clypeorbis mamillata (Schlumb.)	–	–	–	–	–	+	+	–	+	+	+	+
Sulcoperculina sp.	–	–	–	–	–	+	+	–	–	–	–	–
Researchers	G.M. Gasanov 1973–1974				T.A. Mamadov, Sh.A. Babaev, 1974			G.A. Aliev, G.I. Aliev, 1973	T.Ab. Gasanov, 1983–1992			

G. charchaputensis Alijulla, and *Heterohelix striata* (Ehr.) in association with echinoids *Conulus campanaeformis* Melikov et Endelman, *Galerites vulgaris minor* Schuiz, *Echinocorys pyramidata* Portl., *Cardiaster granulosus* Goldf., *Galeaster sumbaricus* Posl., *Seunaster altus* Seunes, and *Homoeaster tunetanus* Pomel., which are of the early Maastrichtian age as well according to conclusion of O.G. Melikov. The unit is 32 m thick.

Upper Maastrichtian (Dashbashi Formation). (4) Alternating white and medium-bedded chalklike, gravely, and clastic limestones; chalklike varieties yield the late Maastrichtian assemblage of echinoids *Conulus magnificus* d'Orb., *Echinocorys arnoudi* Cotteau, *E. perconica* (Hag.), *E. cyplyensis* Lamb., *Coraster vilanovae* Cott., and *Cyclaster integer* Seunes (determined by O.G. Melikov). The unit is 18.0 m thick.

(5) Alternating sugary and gravely limestones; limestone pebbles 1–10 cm in diameter are cemented in the latter varieties by gravely matrix that enclose echinoid spines and fragmented inoceram shells. The unit is 10.5 m thick.

(6) Beds of white medium-bedded and clastic limestones intercalated with shale interlayers (5–10 cm); limestones yield remains of late Maastrichtian echinoids *Echinocorys cypliensis* Lamb., *E. charagoulensis* Gongadze, *E. arnoudi* Cott., *Coraster vilanovae* Cott., *Seunaster lamberti* Charles, and *Homoeaster evaristei* Cott. The unit is 25.5 m thick.

(7) Alternating beds of somewhat sandy and chalklike pinkish limestones; shale interlayers occurring in limestines bear the late Maastrichtian assemblage of foraminifers *Abathomphalus mayaroensis* (Bolli), *Globotruncana arca* (Cushm.), *G. postrugosa* Vass., *Heterohelix striata* (Ehr.), *Stensioeina caucasica* (Subb.), *Pseudotextularia varians* (Rzeh.), *Parrella navarroana* (Cushm.), and *Cibicides bembix* (Marss.), all determined by R.M. Alieva. Along the strike, sandy limestones grade into bluish clayey varieties. The unit is 8.5 m thick.

(8) Medium-bedded chalklike limestones with interlayers of bluish shales; the unit is 12.5 m thick.

Lower Paleocene (Sardash Formation). (9) Pinkish chalklike and sugary limestone beds (50–70 cm), which are intercalated with clayey varieties (20–30 cm) in the upper portion; the latter yield echinoid species *Echinocorys renngarteni* Moskvin and *Homoeaster abichi* Anth., both determined by O.G. Melikov. The unit is 5.0 m thick.

(10) Rhythmically alternating beds of sugary (60–70 cm) and clayey (30–40 cm) limestones; the early Paleocene foraminifers *Subbotina trivialis* Subb., *S. triloculinoides* (Plummer), *Eponides saginaris* N. Byk., and *Globanomalina ehrenbergi* (Bolli) from shale interlayers between limestone beds were determined by R.M. Aliev. The unit is 10.5 m thick.

(11) Saccharoidal and gravely limestone beds (30–40 cm) with interlayers of pinkish clayey limestones; near the top, limestones enclose lentils of clayey shales

(1.5 × 0.1 m), which yield the early Paleocene foraminifers *Subbotina triviaiis* Subb., *S. triloculinoides* (Plummer), *Eponides saginaris* N. Byk., and *Globoanomalina ehrenbergi* (Bolli). The unit is 4.5 m thick.

In total, the lower and upper Maastrichtian substages are 57 and 75 m thick, respectively, and the lower Paleocene deposits are 20 m thick.

In the Gazakh–Agdzhakend depression, the Maastrichtian limestones are locally preserved from erosion in frontal elevations of the Lesser Caucasus piedmonts (Mts. Kyakil' and Malachal, the Dzhogas River and some other areas). Rengarten was first to establish their age in 1959, and his inference was confirmed later on by Abdullaev (1963). Gasanov who discovered *Coraster* cf. *vilanovae* Cott (determined by O.G. Melikov) in association with *Inoceramus lingua* Goldf. and *In. tegulatus* Hag. (determined by O.B. Aliev), when he investigated the Khanlar, Mikhailovka, Kyamarli, Mt. Babakar, and Mt. Malachal localities, came to the same conclusion.

The lower and upper Maastrichtian substages were distinguished for the first time by Mamezade (1967), and our data support his inferences, The lower Maastrichtian sequence of sandy limestones and marls (102 m) bears echinoids *Seunaster altus* Seunes, *Guettaria rocardi* Cott., and *Cyclaster* sp. (determined by O.G. Melikov), as it is established by observations at the Elensutepe, Uchgel', Agdag, Dzegam Quarry, and Dzhagirchai River localities. The upper Maastrichtian deposits are represented by clayey limestones and marls intercalated with shale and sandstone interlayers (20–65 m). These rocks yield the late Maastrichtian echinoids *Homoeaster evaristei* Cott. and *Cyclaster* cf. *integer* Seunes (determined by O.G. Melikov).

In most sections, the Maastrichtian rocks are transgressively overlapped by clayey limestones of the Danian Stage. Deposits bearing Danian fossils (20–30 m) are distinguished in the Damirchilyar, Alpout, Dashsalakhly, Dzegam, and Kyurakchai River sections, where they are represented by clayey limestones that grade upward into the Montian shales of the lower Paleocene.

Reviewer A.S. Alekseev

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Upper Pleistocene to Holocene Deposits of the Tunka Rift (Southern Baikal Region)

G. F. Ufimtsev*, I. V. Shibanova*, N. V. Kulagina*, I. M. Mashchuk*, A. V. Perevalov, V. P. Rezanova**, T. Vogt***, N. V. Ignatova*, and V. A. Misharina***

**Institute of the Earth's Crust, Siberian Division, Russian Academy of Sciences, ul. Lermontova 128, Irkutsk, 664033 Russia*
***Institute of Geology, Siberian Division, Russian Academy of Sciences, ul. Sakh'yanovoi 6a, Ulan-Ude, 670047, Russia*
****Louis Pasteur University, Strasbourg, France*

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Abstract—The Tunka rift valley extending from the SW termination of Lake Baikal by 200 km westward represents a series of large and smaller troughs separated by subordinate elevations. Inversely uplifted bedrock exposures exhibit the Neogene to Quaternary basalts, Pliocene pebblestones, and poorly lithified conglomeratic breccias. Large depressions are filled with sandy, boulder–pebbly, and covering loess deposits. Relationships between these lithological complexes and results of thermoluminescence dating show that they accumulated concurrently during the Pleistocene–Holocene, beginning since 70000 years ago. Alluvial deposits bearing characteristic mollusks are dominant in the sandy complex. These data disprove the idea of a large Tunka Lake (Baikal bay) existence during the Pleistocene time. Proluvial and glacial to fluvioglacial deposits are most characteristic of the boulder–pebbly complex. Eolian processes led to accumulation of loess cover and relevant slope deposits. Alluvial terrace deposits accumulated at various distinct levels of the Irkut River and confluent valleys during the last 40000 years.

Key words: Baikal rift zone, Tunka rift valley, Cenozoic, upper Pleistocene, Holocene, lithological complex, terrace level.

INTRODUCTION

The Tunka rift valley extending from the SW termination of Lake Baikal by 200 km westward represents a system of intermontane troughs separated by transverse and longitudinal horsts or interbasinal barriers (Florensov, 1960). The Tunka depression over 30 km wide is central one in this system. Cenozoic volcanogenic–sedimentary deposits are over 2000 m thick here. In the interrelated Tory, Bystrinka, Mondy, Khoitogol, and Turan depressions, the basement structure and thickness of Cenozoic deposits are known from geophysical data (*Seismotektonika i seismichnost'...*, 1975). The upper part of Cenozoic deposits that is accessible for observations in the rift valley has been formerly studied by various research groups (Adamenko *et al.*, 1983; Kul'chitskii, 1985). Many important data were obtained more than 30 years ago (*Antropogenovye otlozheniya...*, 1964), when E.I. Ravskii and his colleagues outlined principal features of Anthropogene geology in southeastern Siberia.

Recently, we studied a series of sections of loose deposits at the valley floor and in surrounding mountains. We also collected and identified malacofauna, investigated palynological characteristics, and dated sediments using the radiocarbon and thermoluminescence methods. The results obtained are presented in this work. Studying new sections, we did not revise those, which have been investigated in detail before, like, for instance, the Belyi Yar section (Kul'chitskii *et al.*, 1994; Trofimov *et al.*, 1995; *Antropogenovye otlozheniya...*, 1964). Accordingly, we do not discuss here the general problems of the regional Quaternary stratigraphy. This is an objective of future works.

LOOSE DEPOSITS OF THE TUNKA RIFT

Within the rift valley contours, rocks of metamorphic basement and Neogene basalts are exposed in positive relief forms isolating the valley floor areas. Par example, such forms are represented by the Bystrinka and Mondy horsts, and by inversely uplifted blocks on the western slope of the Elovka Spur, where the Pliocene deposits are exposed. As for the largest depressions, their floor is covered everywhere by loose Anthropogene deposits occurring in diverse geomorphological settings. Deposits of this kind occur not only in terrace systems of main stream valleys, but also in the gently inclined piedmont plains, the most remarkable among which is a wide plain adjacent to the Tunka Bold Peaks. In the western floor area of the Mondy depression and around the Khoitogol trough, the Pleistocene morains form a characteristic knob-and-kettle or ridged topography. Eolian sands and loess occur in

ridgelike forms or as a cover above the valley floor and piedmont plains.

The diverse occurrence mode of loose deposits shows that geomorphological criteria of their subdivision and correlation cannot be taken for granted. According to lithological features, deposits under consideration can be grouped into sandy, boulder–pebbly, and cover complexes, the latter composed of loess-like sediments. Below, we characterize these three complexes.

SANDY COMPLEX

Sands cover the greater central areas of the Tory and Tunka depressions and their southern, gently inclined piedmont plains. They often extend into the interfluve areas of northern slope of the Khamar-Daban Ridge up to the relative altitude of 100 m. The same situation is typical of southern margins of the Bystrinka and Turan depression. The domelike Badar massif in central part of the Tunka depression is completely composed of sand and overtops by 150 m the surrounding plain areas of recent subsidence (Florensov, 1960; Ufimtsev, 1992). At the left bank of the Irkut River, the southern side of this massif is truncated by the erosion bench 50–60 m high. The 1.5-m-thick bed of pale-yellow loess-like loamy sand overlies a uniform sequence of medium-grained well-washed sand enclosing rare intercalations and lentils (1–3 cm thick) of silty sand. Sands show the horizontal wavy or vague cross lamination. The sequence is surprisingly uniform and lacking psephite fraction throughout the whole exposure. In these clean sands, pelitic to silty fraction hardly exceeds 10% in abundance. Predominant in sands are grains of 0.25 to 0.1 mm in size (up to 50%).

The middle part of the Badar section yields well-preserved molluscan shells of *Vallonia* ex gr. *pulchella* (Mull.), *Gastrocopta theeli* (West.), *Vertigo (Istmia) modesta* (Say), *V. (I.) alpestris* Ald., and *Euconulus fulvus* (Mull.). These were terrestrial mollusks buried in subaerial conditions. Climate during their lifetime was warmer than now, and average annual temperatures were positive. Age determination by thermoluminescence method showed that upper sands of the section are 65200 ± 400 years old (determination at the Institute of Geology, Siberian Division RAS, Ulan-Ude).

Another 30-m-high bench (Kyren Airport site) is located 6–7 km southwestward of the Badar site at the right side of the Irkut River in the terrace-like structure near the foot of the Khamar-Daban Ridge. The sequence of medium- to fine-grained sands is homogeneous here again. On the basis of slight compositional variations and abundance of pelitic to silty fraction, we tentatively divided the sequence in two parts. The lower part is composed of fine-grained to silty sands displaying the fine wavy or horizontal lamination with rhythms few centimeters thick. Well-washed fine-grained sands above it are light gray, enclosing lenticular intercalations of coarse-grained to gravelly deposits. In addition to horizontal wavy and parallel lamination, we distinguished two types of cross lamination in this part of the sequence. One of them corresponds to the normal fluvial cross lamination showing concentration of coarse material at the base of sand layers up to 0.7 m thick and in basal portions of laminae. It is characteristic of coarse-grained and gravely sands. In the upper part of beds, it is accompanied by swirling textural patterns. Another type of oblique lamination is typical of fine-grained and silty sand beds up to 1.5–2.0 m thick and randomly thicker. Fine parallel laminae are uniformly inclined 15° SE or W in these beds. Laminae are occasionally swirled (turbulence texture), enclosing rounded inclusions (up to 5 cm across) of olive clayey sand. These features and signs of graded bedding exclude a possibility of eolian origin of sands. Their alluvial genesis is also evident from composition of well-preserved fauna of small mollusks found at the depth level of 13 m. We identified there species *Aplexa hypnorum* (L.), *Succinella* ex gr. *oblonga* (Drap.), *Succinea putris* (L.), *Vallonia tenuilabris* (Al. Br.), *Pupilla* ex gr. *muscorum* (L.), and *Vertigo (I.) modesta.* At the higher level of 5–6 m, we found *Succinella* ex gr. *oblonga, Vertigo alpestris, V. modesta, Columella columella* (G. Martens), *Pupilla* ex gr. *sterrii* (Voith), and *Vallonia tenuilabris.* All these species, except for *Aplexa hypnorum*, the dweller of temporary reservoirs, are current habitants of the Baikal region. Their former habitat environments and burial conditions were subaerial. In general, the above malacofauna is of a relatively cryophilic type. Sands from the depth level of 11 m are 31500 ± 2300 years old according to results of thermoluminescence dating. At the depth levels of 8, 12, and 30 m, we discovered scarce diatom remains, including those of extinct Neogene forms (redeposited at the depth of 8 m) and recent freshwater species: epibiontic *Eunotia veneris* (Kutz.) O. Mull. (the depth of 12 m) and littoral planktonic *Cyclotella ocellata* Pant.

The sand sequence of the Kyren Airport site exhibits frequent microfaults dislocating original bedding. We distinguished two dislocation types: (1) step faults with net slip of 10 to 20 cm, which are spaced apart by about 50 cm, and (2) Y-shaped faults, in which the wedge-like microblocks are displaced by about 10 cm downward (Fig. 1).

The Shabartui section is situated at the right side of the Irkut River in the Elovka Spur separating the Tunka and Tory depressions. Sands are exposed here at the top of a high hummocky bench representing the right side of that spur (altitude of 810–820 m) and on its left slope, where they are widespread all along the right side of the Sasaltui Creek valley and have the dead-end contact with its left (western) wall that is composed of crystalline rocks. The spur height above the Irkut River bed is more than 100 m. In a road cut at the top, the exposed sand sequence is over 16 m thick, consisting of two horizons. The 7-m-thick upper horizon is composed of sands, silty in places and displaying horizontal

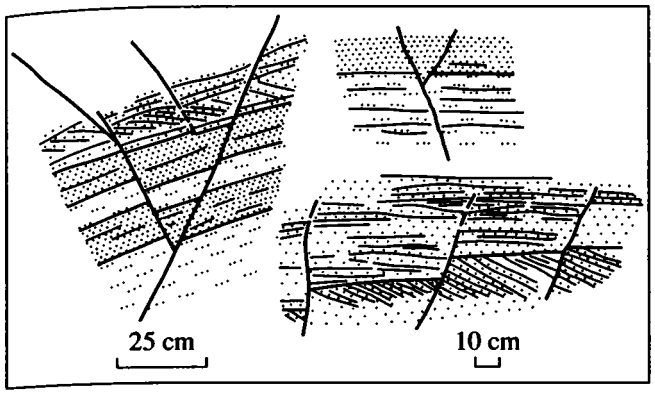

Fig. 1. Microfaults in sands of the Kyren Airport section, southern margin of the Tunka depression.

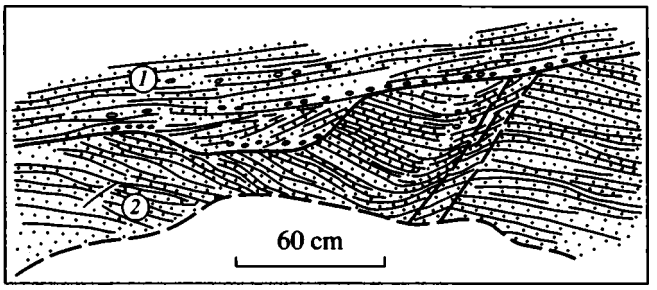

Fig. 2. Infilling of small incisions below the base of upper sand horizon (*1*) of the Shabartui section and patterns of slumping recorded in the lower horizon (*2*).

wavy or parallel lamination. As a rule, laminae are less than 10 cm thick, but some of them can be twice thicker. Pseudomorphs after frost-crack polygons penetrate by 1.5 m downward from the horizon base.

The lower horizon of yellowish to brownish gray fine-grained sands encloses gravelly laminae enriched in limonite. Its typical features are the clinoform surfaces and the oblique parallel lamination dipping 20°–25° SW. In the upper portion, there are scour structures up to 1 m deep. Sands infilling them are either cross-laminated, bearing pebbles at the base of sedimentary rhythms (Fig. 2), or showing oblique lamination so that laminae are conformable to scour boundaries, where minor slumping patterns can be detected as well. Sands from the 10-m-deep level of this horizon are 50400 ± 3000 years old according to results of thermoluminescence dating. It is likely that infilling of old scours exemplifies the oblique lamination in clinoforms, which originated on the sloping parts of past meanders, when the latter migrated across the general stream direction in the ancient Irkut valley.

In the road cut at the western spur slope, sands show the oblique wavy lamination dipping 9° W. Particular laminae are traceable here along the distance of hundreds meters and thus cannot represent elements of meander clinoforms, because we must assume in such a case that the ancient stream bed was as deep as dozens of meters. Another possibility is to assume the eolian genesis of sands, the more so that the present-day eolation is active here, and the road-cut wall is decorated with eolian ripple marks. It is necessary to take into account, however, that sands with oblique lamination at the right side of the Sasaltui Creek valley yield fragmented shells of large pulmonate mollusks *Anisus (Gyraulus) acronicus* (Fer.), the habitants of permanent shallow reservoirs, and *Pupilla* sp. The laminae inclined westward reflect most likely a consequence of recent uplift, in response to which the right-side bench of the Elovka Spur was tilted to be in concordance with the oblique horst on the spur left side. One more peculiarity that is observed at the spur top is occurrence of gravel lentils (up to 30 cm thick) and angular 20- to 30-cm-wide granitoid boulders (erratic material?) at the base of the lower sandy horizon.

Palynological spectra characterizing sands of the Shabartui section are impoverished, consisting of recent and redeposited Miocene miospores. Excluding the latter, we may assume that sands accumulated during a cold climatic phase. Single frustules of diatom *Alveolophora areolata* (Moiss.) Moiss., comb. nov., the extinct planktonic species of the Miocene time, were also found here.

The Tibel'ti section is located in the eponymous river basin, the southeastern corner of the Tory depression. Above the boulder–pebbly deposits, there is exposed the 10-m-thick sequence of sands with horizontal and fluvial cross lamination. Distinguishable sedimentary cycles 5–7 cm thick consist of inaequigranular and fine-grained sands. Small scours up to 1 m deep are recorded in the upper part of this section. The depth level of 7.2 m from the top, where we observed pseudomorphs after ice wedges, is likely an analogue of the level of ice-crack polygons below the upper sandy horizon of the Shabartui section. In the Aerkhan section (the Irkut River right side in western area of the Mondy depression), a thin sandy bed with involution structure overlies the boulder–pebbly deposits of the Pleistocene moraine (*Antropogenovye otlozheniya...*, 1964; *Nagor'ya Pribaikal'ya...*, 1974). Another bed or lentil up to 3.5 m thick is localized inside morainal boulder–pebbly deposits, and its base and bedding planes dip 240° SW under angle of 20°. Alternating sand strata up to 0.5 m thick show two types of structural patterns: parallel and cross-ripple-mark lamination. The latter, representing a system of S-shaped lenticular laminae up to 20 cm long, is similar to that of loess-like loamy sands and resembles the present-day eolian ripple marks. At the same time, the sands in question are found inside morainal deposits, and we should assume that their body is an esker of fluvioglacial origin. The heavy fraction of sands is rich in pyrite, and this suggests a local provenance of terrigenous material (see below). The underlying pebblestone is well washed and of the same origin.

Eolian deposits widespread in many areas of the Tunka rift (Ufimtsev *et al.*, in press) represent a particular component of the sandy complex. Principal covers of eolian sands are situated on western slopes of the Elovka and Nilovka spurs; they also envelope the Badar sandy massif. These products of eolian reworking of underlying sands show characteristic microrelief of the knob-and-kettle type. Morphoscopy showed that quartz grains are polished by river streams, and that distance of their transportation was insignificant. In the eastern area of the Khoitogol depression, eolian sands rest on the surface of the 14-m-high terrace, the upper layer of which is 29300 ± 1000 years old according to results of thermoluminescence dating. The Mt. Khairkhan at this site (relative elevation 150 m) is also composed of eolian sands.

Seven age determinations have been obtained for the first time for eolian sands of the Tunka valley. Thermoluminescence and IRSL analysis were carried out at the Institute of Geography, Cologne University of Germany (Ufimtsev *et al.*, 1999). The main phase of eolian sand accumulation lasted from 16120 ± 3580 (IRSL 16120 ± 2730) to 8310 ± 2630 (IRSL 8360 ± 1360) years ago.

Compositional peculiarities of sands (eolian varieties excluded) are illustrated in Fig. 3. The diagram shows that they mostly consist of quartz and feldspar grains, the former slightly more abundant. The heavy fraction of sands from the Aerkhan section is extraordinary rich in pyrite grains (up to 43%!). This suggests their proximity to provenance, where some mineral deposit was located. Sands from other sections, except the Tibel'ti site, yield heavy fractions of magnetite-ilmenite composition with comparable proportions of stable and unstable minerals. Stable minerals are relatively more abundant at the Badar and Shabartui sites, and others slightly prevail in sands of the Aerkhan and Tibel'ti sites. We suspect that this is a consequence of clastic material transportation over different distances.

Boulder–Pebbly Complex

Two age groups of boulder–pebbly deposits have been distinguished in the Tunka rift. One includes boulder and boulder–pebbly deposits of the Anosovka Formation and the so-called Mondy Conglomerate (Adamenko *et al.*, 1983; *Litologiya tretichnykh...*, 1972). These occur on sides of small ridges separating one depression from another and in minor Bystrinka and Mondy depressions superimposed on those ridges, which are located near the junctions zone of the Tunka, Baikal, and Khubsugul rifts. In the Bystrinka and Mondy depressions, deposits of the Anosovka Formation yield bones of Pliocene small mammals (Adamenko *et al.*, 1983) and are exposed in inversely uplifted blocks. In the central area of the Bystrinka depression (Il'cha section and others), the Anosovka Formation is composed of poorly lithified conglomerates (bodied pebblestone with small boulders). Many pebbles are subangular and broken. Toward the depression walls, pebblestone grades into conglomeratic breccia and slope gravel with large floats of weathered metamorphic schists. Pebblestone bodies and individual pebbles are fractured (Ufimtsev, 1992).

Deposits of the Anosovka Formation are distinctly tectonized in the Bor'skii Irkut section near the northern wall of the Turan depression. Proluvial deposits and talus are composed here of fragments of Neogene basalts derived from the Nilovka Spur and rest within a recent fault zone. Rock fragments are weathered to the depth of 2 cm, and their surface is slickesided. Many large fragments are tectonically fractured (Ufimtsev, 1996).

Light mineral fractions separated from the Il'cha and Bor'skii Irkut sections include aggregates of secondary minerals that originated by weathering. In heavy fractions from the Il'cha section, ilmenite and titanomagnetite are more abundant than magnetite. The latter can be even missing from their composition characterized by a high proportion of stable minerals (Fig. 3). In heavy fractions from the Bor'skii Irkut section, abundance of magnetite is highly variable, because proluvium accumulated here at the expense of basaltic flows disintegration.

Boulder pebblestones of younger ages are characteristic of coalescing talus fans at the base of northern fault scarp of the rift valley. They characterize as well the Pleistocene morainal and fluvioglacial deposits especially typical of the Mondy depression. Similar deposits are also known at foot slopes of minor ridges separating depressions. Large bodies of boulder pebblestone are embedded in sands within junction areas of rift valley with gorges of the Khamar-Daban Ridge. An example of this kind is the Tibel'ti section.

Pebbles and boulders up to few cubic meters in size, which occur in the upper part of proluvium at the foot of the Tunka fault scarp, are of the Holocene age. At any rate, sand and gravel, the basal facies of the mountain apron sequence, yield remains of terrestrial mollusks *Succinea putris* and are less than 3000 years old at the depth of 4 m, as their age is below the determination limit of thermoluminescence analysis.

In the marginal part of the Tunka depression, boulder pebblestones constitute a bajada adjacent to the Elovka Spur foot, the terraced continuation of which is plunging toward the depression center under lacustrine-bog sediments accumulated within a zone of intense recent subsidence. The quarry at the northeastern side of the Tunka Village exhibits to the depth of 8–10 m the homogeneous pebblestone sequence enclosing rare boulders and gravel lentils up to 0.5 m thick. Lamination in the sequence is horizontal, slightly wavy. According to the date obtained by radiothermoluminescence method, the gravel lentil at the depth of 4.5 m is 70000 ± 10000 years old. In other quarry located northward near the road turn to the Saganur village, the same sequence of boulder pebblestones is buried under the

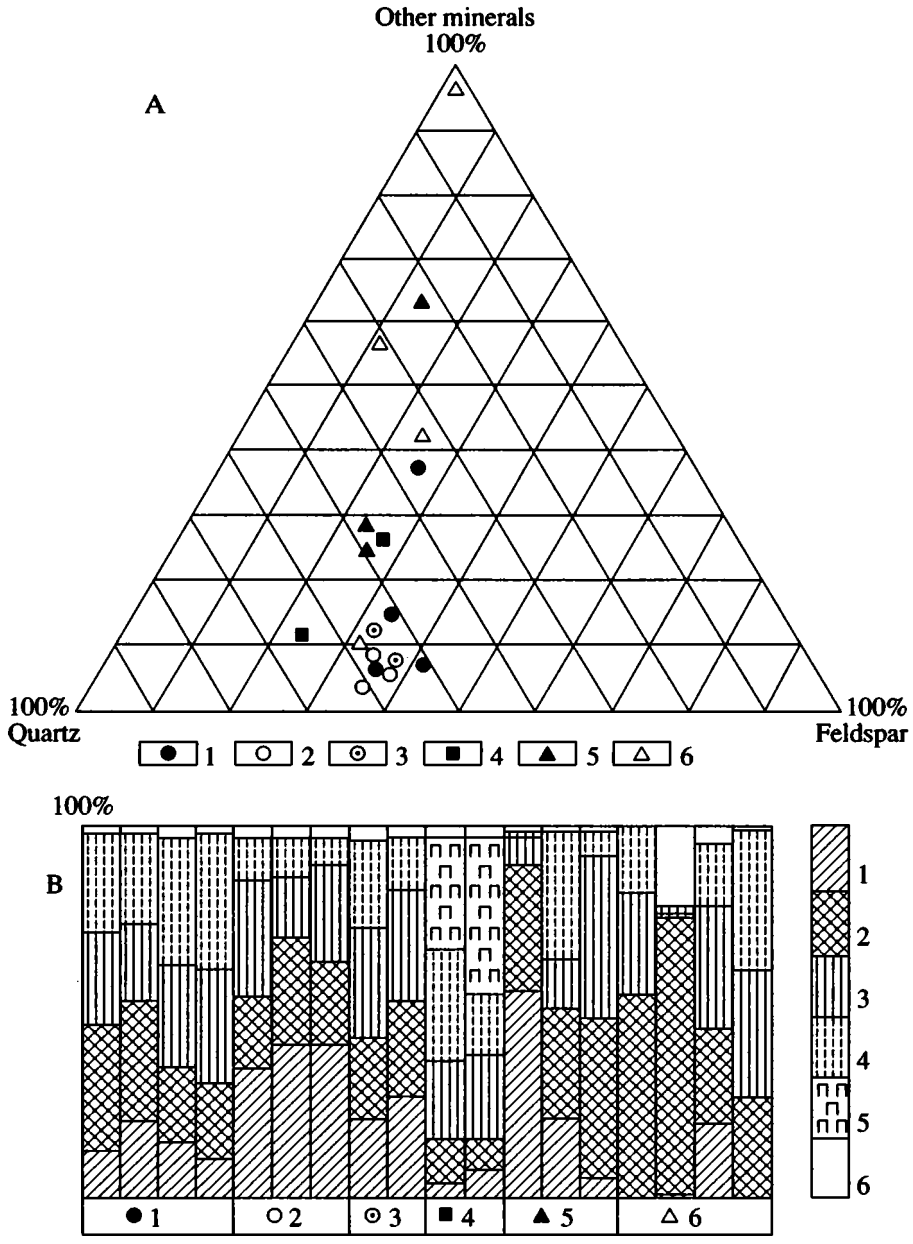

Fig. 3. Mineral compositions of light (A) and heavy (B) sand fractions from the Tibel'ti (1), Badar (2), Shabartui (3), Aerkhan (4), Bor'skii Irkut (5), and Il'cha (6) sections. Right column in diagram (B) shows symbols of magnetite (1), ilmenite (2), stable minerals (3), unstable minerals (4), pyrite (5), and others (6).

cover of loamy sands that are 1.5 m thick and enclose a fossil soil horizon. Below that cover, there is a lentil of gravely sands, which turned out to be 39000 ± 6000 years old at the depth of 1.75 m (a result of radiothermoluminescence dating).

In the eastern part of the Tunka depression, the same deposits overlie basal portions of the Tala cinder cones (Florensov and Loskutova, 1953). The above IRSL dates thus indicate an earlier age of recent volcanic eruptions.

Deposits of presumable glacial origin are widespread in the east of the Mondy depression. Their distribution areas are characterized by the knob-and-kettle topography. Morainal ridges up to 50 m high are also known on uplifted faulted blocks in the northern depression area. To the east of the Mondy depression, within the Kharadaban tectonic block slotted by the narrow antecedent valley of the Irkut River, similar boulder–pebbly deposits infill incisions, which are up to 60 and more meters deep. In a road cut at the left side of the river near the Khu-

lugaim Creek mouth, these deposits are exposed down to the depth of 30 m. Gray boulder–pebbly facies of this sequence enclose some rock fragments up to two cubic meters in volume and vaguely outlined lentils of sandy–silty material up to 2 m thick. These lentils show alternation of 10- to 20-cm-thick beds composed of gray clean sand, small-sized gravel, olive-gray silty sand, and fine-laminated siltstone. The compact, poorly sorted clastic matrix cementing boulders and pebbles is rich in silty–pelitic material (glacial flour?). In the eastern corner of exposure, the tilted boulder–pebbly beds truncate beds of the same composition, which are located below and westward. In the easternmost part, they are in contact with a horizontally bedded member.

The described deposits cannot be reliably dated, and their morainal genesis is under question. In contrast to morains known in the Irkut riverhead area and Oka Plateau, where boulders and pebbles often show glacial striation, the material of Mondy moraine (?) is lacking such marks. It is also remarkable that the Irkut River valley between the Tunka Bold Peaks and Munku-Sardyk Ridge does not resemble in morphology a glacial trough. Thalwegs connecting the Munku-Sardyk glacial cirques with the Irkut valley are as steep as interfluve slopes, and this is favorable for development of surging but not normal glaciers. In general, there is a lot of unsolved problems concerning Pleistocene glaciation in the study region.

Cover Complex

Covers of light (yellowish) gray loess-like loamy sands and loams envelope various relief forms composed of crystalline and loose deposits. These monotonous sediments can be more than 5 m thick, and their sections are well exposed in road cuts along the Kultuk–Mondy highway, when it crosscuts ridges separating depressions. One section of yellowish gray loamy sands with columnar jointing was studied in basal portions of gently inclined glades at the foot of the Khamar-Daban Ridge (eastern surroundings of the Zaktui Village, the Irkut River right bank). The apparent thickness of sediments is more than 5 m, and they all show wavy-lenticular lamination. The S-shaped lenticular laminae mimic manifold the eolian ripple marks on adjacent vegetation-free surface areas and leave no doubt about the eolian origin of these loess-like sediments. In the Nilovka Spur between the Malaya and Bol'shaya Zangisany valleys, the cover of loamy sands overlies boulder–pebbly deposits of presumable fluvioglacial origin. Microrelief above the cover is of the knob-and-kettle type. It was presumably formed by the permafrost degradation, in particular, by thawing of ice wedges.

Eolian material appears to be an important component of recent sedimentation in the Tunka depression. In inner depression areas, it seems to be mixed with main mass of other sedimentary materials, gaining significance in areas of subaerial sedimentation, e.g., on gentle slopes of the Elovka Spur and in glades of synonymous river valley. In ravines of this locality, brown to yellowish gray loess-like loamy sands and loams are exposed to the depth of 5 m. These sediments enclose isolated boulders and lenticular waste bodies up to 10–20 cm thick. The lentils dip toward the slope line, whereas vague lamination in loams and sands is approximately parallel to the latter. A dismembered fossil soil horizon 10–15 cm thick with tiny carbonaceous patches at the base is buried here at the depth of 0.5–1.0 m.

Slope loamy sands of the Elovka valley yield abundant remains of mollusks *Succinella* ex gr. *oblonga, Succinea putris, S.* sp., *Cochlicopa lubrica* (Mull.), *C. lubricella* (Porro), *Columella columella, C. edentula* (Drap.), *Vallonia costate* (Mull.), *V. tenuilabris, Pupilla* ex gr. *muscorum, P. steerri, P.* sp., *Vertigo (I.) modesta, V. (I.) alpestris, Nesovitrea hammonis* (Strom.), *Euconulus fulvus, Gastrocopta* ex gr. *theeli, Discus ruderatus* (Stud.), and *Bradybaena schrencki* (Mull.). Remains are well preserved, and all the distinguished taxa are known among present-day mollusks of the Tunka depression. The malacofauna was buried under subaerial conditions, and its habitat environments did not differ from those observable now.

It is likely that the loess-like cover and slope deposits accumulated concurrently with eolian sands. The latter were deposited after a short-distance transportation of alluvial and lacustrine sands, and the former represent accumulations of dust transported by wind over long distances. At the Tibel'ti site of the Tory depression, loamy sand covers above alluvium of the first and second stream terraces of the Irkut valley are 5180 ± 40 and 9280 ± 40 years old, respectively, according to results of radiocarbon dating (*Putevoditel'...*, 1982). Bones found by Yu.V. Ryzhov in slope deposits of one of the Elovka ravines were identified by N.P. Kalmykov as remains of *Eqvus* sp. and *Bison pariscus*. The inferable age of these remains may range from the late middle to late Pleistocene.

DEPOSITS OF STREAM TERRACES

In valleys of the Irkut River and Ikhe-Ukhguni confluent crossing the Tunka depression, there are three distinct terrace levels: a high flood plain and first and second terraces above it. Sections of these geomorphological elements are composed of dominant sandy and less abundant boulder–pebbly material.

Stratotypes of first and second terraces were described at the Tibel'ti site of the Tory depression, where age of sediments is established with the help of radiocarbon dating (*Putevoditel'...*, 1982). The second stream terrace is 14 m high here, and sediments from the middle part of its section yielded radiocarbon dates of 40060 ± 820 and 31860 ± 37 years. A close age of 29300 ± 1000 years was measured using thermolumi-

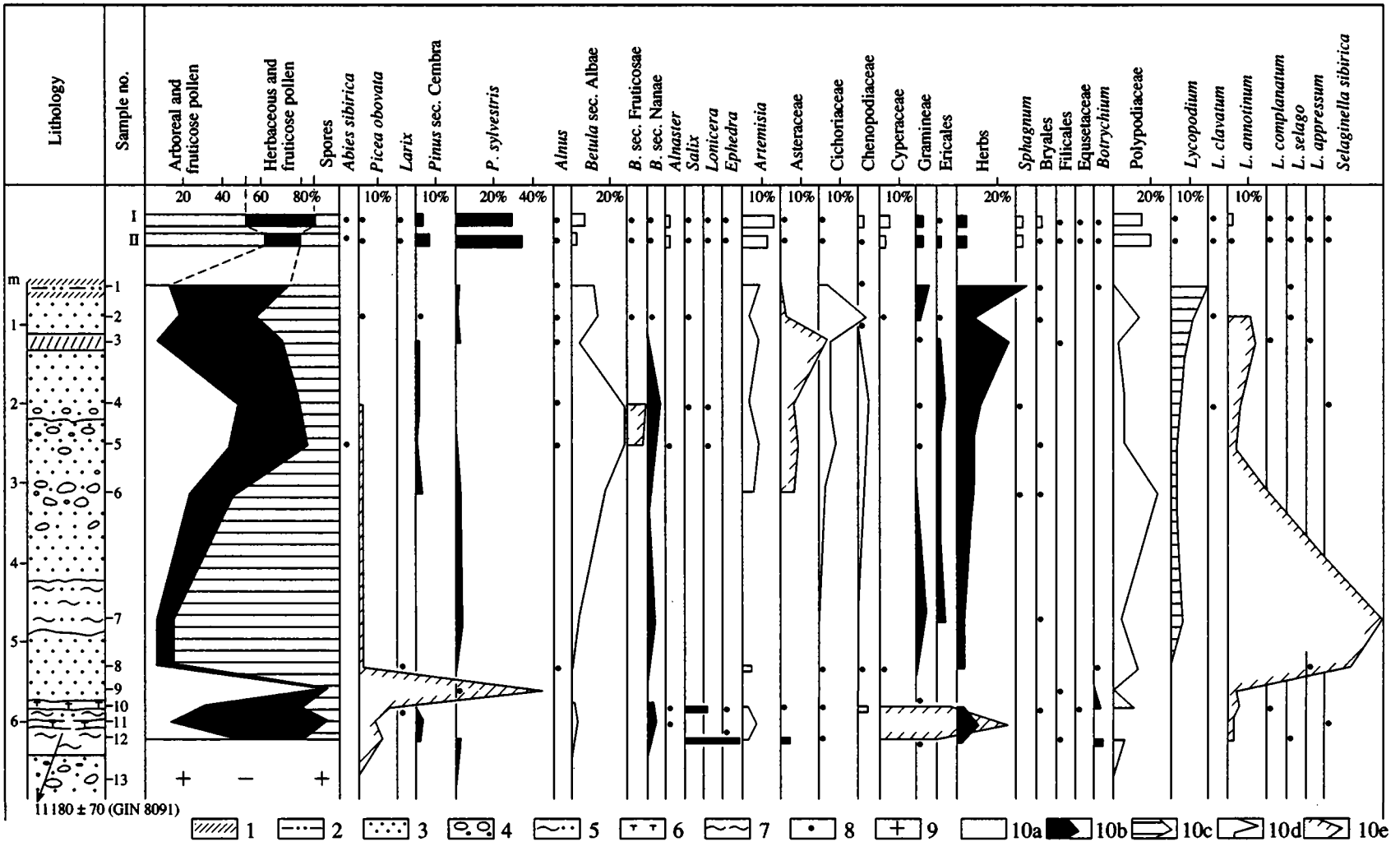

Fig. 4. Palynological diagram for the terrace section at the left side of the Malaya Taitorka River, Shimki Village site in the Irkut right-bank area: (1) buried soil; (2) loamy sand; (3) sand; (4) pebbles and gravel; (5) loam; (6) peat; (7) clay; (8) abundance of palynomorphs less than 1%; (9) single spore and pollen grains; (10a–10e) variation curves for different palynological groups and taxa. Also shown in diagram are average abundances of palynomorphs in the recent alluvium of the Irkut River: (I) right bank near the Turan–Nilova Pustyn' highway bridge; (II) left bank near the Kultuk–Arshan highway bridge.

nescence analysis in the 14-m-high terrace of the Ikhe-Ukhguni valley in the east of the Khoitogol depression, where the terrace section is mostly composed of boulder–pebbly material (some boulder are over two cubic meters in volume), and sands constitute only its upper part.

At the Tibel'ti site, the first terrace above the flood plain is 8–9 m high. Known radiocarbon dates for sediments from the middle interval of its section are 9800 ± 130 and 10300 ± 80 years (*Putevoditel'...*, 1982). Terraces of this level are widespread in the Irkut valley, and the 7- to 8-m-high terrace near the Shimki Village at the right bank of the Malaya Taitorka River (Fig. 4) belongs to this group. The lower part of bank precipice is composed here of fluvial pebblestone with rusty sandy matrix (Sample 13; depth level of 6.7 m). Single pollen grains of spruce and pine in association with fern spores were detected in this rock. Spore–pollen spectra in the overlying member of clay, loam, and peat (samples 9–12; depth interval of 5.6–6.2 m) show a considerable abundance of arboreal and fruticose taxa, especially of spruce. Pollen of *Larix, Pinus sylvestris, Pinus sibirica*, and *Betula* is less abundant. *Salix* pollen is dominant in the fruticose group, and Cyperaceae are main representatives of herbaceous taxa. Such pollen spectra correspond to the forest community of spruce, larch, birch, Scotch and cedar pines. Osiers and grassy communities with Cyperaceae populated damped areas, whereas sagebrush, *Ephedra*, and grape-fern grew in open and dry sites. Frustules of planktonic diatomaceous algae *Cyclostephanos costatus* Log., Lupik. et Chrus. var. *costatus*, which were also found in the indicated interval, represent pelagic dwellers of deadwaters and slow streams. A peat sample from the depth of 5.8 m yielded the radiocarbon date of 11180 ± 70 years (GIN-8091) that suggests its accumulation during a warming phase at the end of the Sartanian time (Trofimov *et al.*, 1995).

Palynological spectra in overlying sand (samples 6–8; depth interval of 5.6–3.0 m) are rich in spores of *Lycopodium annotinum* that grows now in coniferous forests, poplar groves, and brushwood (*Flora Tsentral'noi...*, 1979). As one can judge from low abundance of arboreal taxa, it was a time of the rarefied vegetation cover with widespread plant communities of open landscapes. At the depth level of 3.1 m, we detected benthic diatoms *Hantzschia amphioxys* f. *capitata* O. Mull., *Pinularia borealis* Her. var. *borealis* (North Alpine species typical of northern and mountain reservoirs), *Cymbella ventricosa* Kutz. var. *ventricosa, Fragilaria construens* var. *triundulata* Reich. (epibiontic forms), and *Tetracyclus rupestris* (A. Br.) Grun. (another typical North Alpine species) in association with planktonic *Cyclotella* Kutz. sp. These species point to a cold and humid climate at the end of the Sartanian time.

Pollen spectra from sands and pebblestones of the interval of 3.0–2.0 m (Sample 45) show an increased role of arboreal and fruticose pollen mainly on account of *Betula* sec. *Albae* and *Betula* sec. *Nanae*. In addition to spruce and pine pollen, there is that of fir, alder, and honeysuckle. Herbaceous pollen is diverse. Changes in vegetation indicate some climatic amelioration (early Holocene?).

Abundance of arboreal and fruticose pollen, especially of *Betula* sect. *Albae* and *Betula* sect. *Nanae*, decreases in the depth interval of 2.0–0.5 m (Samples 1–3), where pollen of herbaceous plants is more diverse than below. According to composition of herbaceous pollen group, open landscape areas were populated in the corresponding period by xerophytes and meadow grasses. The mixed forest communities of birch, spruce, pine, and alder trees were scattered in river valleys. In general, this vegetation presumably corresponded to the next (Pit–Igarka) phase of climatic cooling about 9800–9300 years ago, i.e., at the time of the early Holocene cold epoch (Kind, 1974).

As compared to terrace deposits, a much higher content of *Pinus sylvestris* pollen is recorded in palynological spectra characterizing recent alluvium of the Irkut River (average data for 10 samples; see Fig. 4). Abundance of this component began to grow about 6000 years ago (Tomskaya, 1981; Andreev *et al.*, 1992; Bezrukova *et al.*, 1996).

Malacofauna is also abundant in the section under consideration. Species *Cochlicopa* sp., *Pupilla* sp., *Vallonia costate, V. tenuilabris, Discus ruderatus, Nesovitrea hammonis, Euconulus fulvus*, and *Vertigo (I.) modesta* are found in the depth interval of 5.75–2.2 m that is composed of sands and loams. At the depth level of 6.3 m, gray clay yields *Lymnaea (Galba)* sp. (juv.), *L. (Peregriana) ovata* (Drap.), *L. (P.) balthica* (L.), *L. (P.) ampullacea* (Ross.), *Aplexa hypnorum, Succinella* ex gr. *oblonga, Succinea putris*, and *Columella columella*. All the listed mollusks exist at present. Their assemblage from the higher level consists of terrestrial species only, whereas the deeper one is dominated by dwellers of shallow water basins.

One more assemblage of mollusks *Lymnaea (Galba) truncatula* (West.), *L.* sp., *Anisus (Gyraulus) acronicus* (Fer.), *Cochlicopa?* sp., *Gastrocopta* ex gr. *theeli, Pupilla* sp., *Vallonia costate, V. tenuilabris, Punctum pugmaeum* (Drap.), *Discus ruderatus*, and *Nesovitrea* sp. (juv.) is associated with rare ostracodes in the middle interval of gray clean sands, which are exposed in the 5-m-thick terrace section near the Shimki Village at the right bank of the Irkut River. This malacofauna characterizes a phase of climatic warming. Originally, Ravskii suggested the Eopleistocene age of this terrace (see in *Antropogenovye otlozheniya...*, 1964). Later on, he correlated its origin with the Kazantsevoan Interglacial (Ravskii, 1972). Available radiocarbon dates imply however its younger age.

CONCLUSION

Loose upper Pleistocene sediments widespread in depressions of the Tunka rift can be divided into three lithological complexes of sands, boulder–pebbly deposits, and upper cover of loess-like loamy sands. Older than others are boulder and boulder–pebbly accumulations of the upper Pliocene, which are exposed between individual depressions in the inversely uplifted structures with subordinate inner troughs. In main depressions (Tory, Tunka), most representative Pleistocene sections are also confined to recently uplifted areas. For instance, these are sections in southern margins of depressions risen up in the course of the Khamar-Daban Ridge arching (Kyren Airport site), the Badar massif section of an inverted domelike structure, the Shabartui and Khulusansata sections adjacent to barrier strictures between depressions. Central lowland areas of depressions host lake systems and floodplains, the sites of intense recent subsidence (Florensov, 1960; Ufimtsev, 1992).

Available dating results imply that various lithological complexes were concurrently deposited in the second half of the late Pleistocene. This is an obstacle necessitating elaboration of special approach to the study, subdivision, and correlation of loose sedimentary sequences. Lithology of the Upper Cenozoic sections in the Tunka rift, peculiar lamination of sediments, and remains of malacofauna suggest that they characterize alluvial and proluvial facies. Lacustrine deposits of considerable thickness have not been detected in the studied sections. Consequently, the idea that the Tunka rift valley hosted large Pleistocene lakes or represented a bay of the *Great Baikal* appears to be poorly substantiated.

One of the main problems is subdivision of loose deposits in the Tunka and other rift valleys of the Baikal system. It is a lucky chance to obtain representative palynological spectra characterizing sandy and boulder–pebbly deposits. Similarly rare are finds of mammalian remains. Applicability of radiothermoluminescence dating is also limited, especially with respect to sandy deposits, and we used in this work only the most reliable dating results. Advantages and limitations of this method cannot be properly discussed now, as a more representative database is required.

At the same time, fossil mollusks occur in most sections of the Pleistocene sands. Compositional changes in their assemblages and abundance rates of taxa in particular sediments offer a possibility to use these fossils for the "high-resolution" stratigraphy of Pleistocene sediments in the Tunka rift valley.

ACKNOWLEDGMENTS

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Reviewer M.A. Akhmet'ev

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Application of Magnetic Susceptibility Measurements in Paleogeographic Investigations

V. A. Bol'shakov

Moscow State University, Vorob'evy gory, Moscow, 119899 Russia

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Abstract—A series of publications, in which the magnetic susceptibility measurements are used for paleoclimatic interpretations and geological correlation of the loess formation, is analyzed. As is shown, the magnetic susceptibility parameters are of a limited value for the indicated purpose, and their interpretation should be specially substantiated in each concrete case.

Key words: magnetic susceptibility, paleoclimate, correlation, loess, buried soil, oxygen-isotope data.

The magnetic susceptibility parameters (κ) of sedimentary rocks are frequently used in paleogeographic studies, but not all the inferences based on these parameters are correct and convincing. In my previous works (Bol'shakov, 1996, 1997, 2000), I considered methodological and methodical problems of interpreting the rock magnetism, in particular, their magnetic susceptibility (MS) parameters. It was shown that MS of deposits depends on many factors, such as climatic conditions of sedimentation, remoteness of magnetic material provenance, local geochemical and geomorphological conditions, peculiar features of rock lithology, composition of underlying rocks, time, and others. With such a multi-factor dependence, we should establish basic causes responsible for MS variations to understand their implications for particular problems. In other words, it is necessary to know the specific physical mechanism responsible for origin of a magnetic signal, and only after this, one can judge whether the posed problem will be solved using MS measurements. This requirement is a basic one in the methodological approach that allows obtaining of non-controversial results when using methods of rock magnetism and, particularly, the MS measurements (Bol'shakov, 1997, 2000). Unfortunately, when requirements of this methodological approach are inconsistently followed, we obtain paradoxical results. For instance, the spectral analysis of MS parameters carried out by Beget and Hawkins (1989) and by Kukla *et al.* (1990) revealed periodical changes of κ values in loess–soil sequences of Alaska and China, which are comparable with variation cycles of orbital elements. Regretfully, secular changes of sedimentation rate has not been taken into account in the first case, and incorrect physical prerequisites have been used in the second attempt to take this factor into consideration (Bol'shakov, 1997, 2000). To put it another way, time scales in the both cases were primarily inconsistent and incorrect, although the results of the spectral analysis did show that periods of κ variations are correlative with orbital cycles. Accordingly, the important conclusion (correct, in the authors' opinion, but unjustified from my standpoint) has been drawn that orbital variations exerted a decisive effect on loess formation in our planet. My analysis of inferences and conclusions, presented in works mentioned above (Bol'shakov, 2000) and in some others I refer to below, suggests that there is a certain mystic attitude of authors to the possibilities of MS method, although their own empirical data refute universalism of the latter. Such examples of interpreting MS measurements cause the distrust only.

The work by Thouveny *et al.* (1994) is one of few others, in which MS measurements are correctly used in terms of methodology. Examining deposits of volcanic lakes of France, they proposed a simple model of climate-induced κ variations that was supported by data of paleobotanical, granulometric, mineralogical, geochemical, and diatomic analyses. Under cold conditions, freezing–thawing processes in the ground enhanced erosion of volcanic rocks and deposition of clastic material (κ increase). At the time of warming, the accelerated processes of soil and plant formation lead to the intensification of organic sedimentogenesis that increases proportion of nonmagnetic fraction (κ decrease). With the help of radiocarbon dating and methods mentioned above, paleoclimatic variations during the last 140 ka were comprehensively studied, and a "fine structure" of climatic changes during the oxygen-isotope (OI) stage 5*e* of the last interglacial period was confirmed. Thouveny *et al.* (1994, p. 503) pointed out: "The magnetic susceptibility is thus strongly sensitive to variations of the local climate and constitutes an accurate proxy record, particularly useful when paleobiological or organic geochemical information loses some degree of significance (that is, under the coldest conditions)."

The aim of this work is to analyze the concrete examples of using MS measurements as a tool of geological correlation and other works related to different aspects of paleoclimatic studies. The analysis is dedicated to rocks of the loess formation (LF) in different regions of the world, because the MS method was most widely used to study these rocks.

Before the analysis of publications, I should note that the MS measurements are especially effective when κ values exceed 10^{-4} SI units. Such values usually imply presence of ferrimagnetic minerals (magnetite, maghemite, greigite or titanomagnetite) in sediments. If κ values are lower, the interpretation of MS measurements is difficult, requiring additional studies. A good illustration is the work by Hounslow and Maher (1999) who showed, using data of the X-ray study, Mössbauer spectroscopy, electronic and optic microscopy, and sediment dispersion, that κ values lower than 5×10^{-5} SI units are ineffective for acquiring paleoclimatic information from MS measurement results. This example also shows that low magnetic susceptibility of deposits deprives the MS method of its advantages, which are the insignificant time consumption, low cost, and possibility to obtain data without destruction of studied subjects, i.e., to carry out measurements directly in a natural exposure or in a core section of bottom sediments. One more advantage of MS measurements is possibility to obtain data in a digital form suitable for a direct processing and further interpretation.

CRITICAL ANALYSIS OF SOME RESULTS BASED ON MAGNETIC SUSCEPTIBILITY MEASUREMENTS

Geological correlation of Pleistocene deposits and interrelated climatic interpretations are usually based on comparison of MS parameters and OI data. Without this, the MS measurements are much less effective, as the widely used OI time scale is climatochronostratigraphic in essence.

The effect of climatic conditions on the magnetic susceptibility of soils has long been known (Le Borgne, 1960; Babanin, 1971; Vadyunina and Babanin, 1972; Tite and Linington, 1975; Mullins, 1977; Smirnov, 1978). Paleoclimatic aspects of magnetic susceptibility measurements in loess and soil were also touched upon in these works. In due time, Mullins (1977) warned against hasty attempts to use the κ values of soils (or ratio between these parameters in soils and loess κ_{bs}/κ_l) to solve the reverse problem, i.e., to determine paleoclimatic conditions during the formation time of LF horizons based on results of MS measurements. His was aware of multifactor dependence of κ values on various, not only climatic, conditions (for instance, on composition of underlying rocks), as is mentioned above.

Among the early and well-known publications, there are works by Heller and Liu (1984, 1986), in which the MS variations in LF horizons of China (the Luochuan section) were attributed to influence of paleoclimatic factors. They concluded that magnetic susceptibility could be indicative of paleoclimatic changes based on a regular variation of κ values throughout the section: κ values are higher in soils, formation of which is related to interglacial stages, and lower in loess horizons deposited during glaciation stages. When Heller and Liu correlated MS parameters and OI data, they found them rather consistent with each other. The consistency is, however, not absolute, not "one-to-one" as mentioned in their work published in 1986. The main apparent inconsistency is that the Matuyama–Brunhes (M/B) reversal recorded in the OI Stage 19 in deep-sea sediments is at the level 21st to 23rd excursions in the Luochuan section, which corresponds to the buried soil 8 (BS 8). Admitting the chemical nature of characteristic rock magnetization in this section, Heller and Liu (1986) shift the B/M reversal boundary to the loess bed about 1 m above the BS 8 roof, whereas in earlier publications (Heller and Liu, 1984; Liu *et al.*, 1984) it was established inside and even at the base of BS 8. Displacement of inversion records during secondary magnetization is a well-known phenomenon (Bol'shakov, 1995, 1996), although the displacement magnitude should be assessed in a special manner (Bol'shakov, 1999). The displacement in question is voluntary and insufficiently justified, because loess, in the authors' opinion, accumulated under cold arid conditions, whereas the B/M reversal is established in the warm OI Stage19.

Heller and Liu (1986) also attempted to carry out the spectral analysis of MS parameters. They accepted the time scale based on paleomagnetic time markers and the constant sedimentation rate throughout the section, but did not obtain a significant result. Tuning the κ values to insolation curves estimated by Berger (1978), they revealed however periodical changes approximating cycles of 40 k.y., which characterize variations of the equator plane obliquity to the ecliptic plane. Heller and Liu (1986) concluded that cycles of 23 k.y. are missing because of a low effect of precession-related seasonal variations of insolation on the LF origin in China. This inference is doubtful, as the territory of China is under influence monsoons (paleomonsoons) (Banerjee *et al.*, 1993), the intensity of which is interrelated with precession of the earth's axis (Rossignol-Strick, 1983; Prell and Kutzbach, 1987). It is even stranger that periods of 100 k.y. of κ variations, which are clearly seen in Fig. 1 borrowed from the work by Kukla *et al.* (1988), have been missing. I did not plot MS data for the Luochuan section reported by Heller and Liu (1984, 1986), because they used a logarithmic scale that conceals some details, and because the both are co-authors of the work by Kukla *et al.* (1988). The mentioned unsatisfactory results of the spectral analysis are likely a consequence of insufficiently elaborated time scale that was applied to the Luochuan section. In addition, the effect of various factors on the κ value

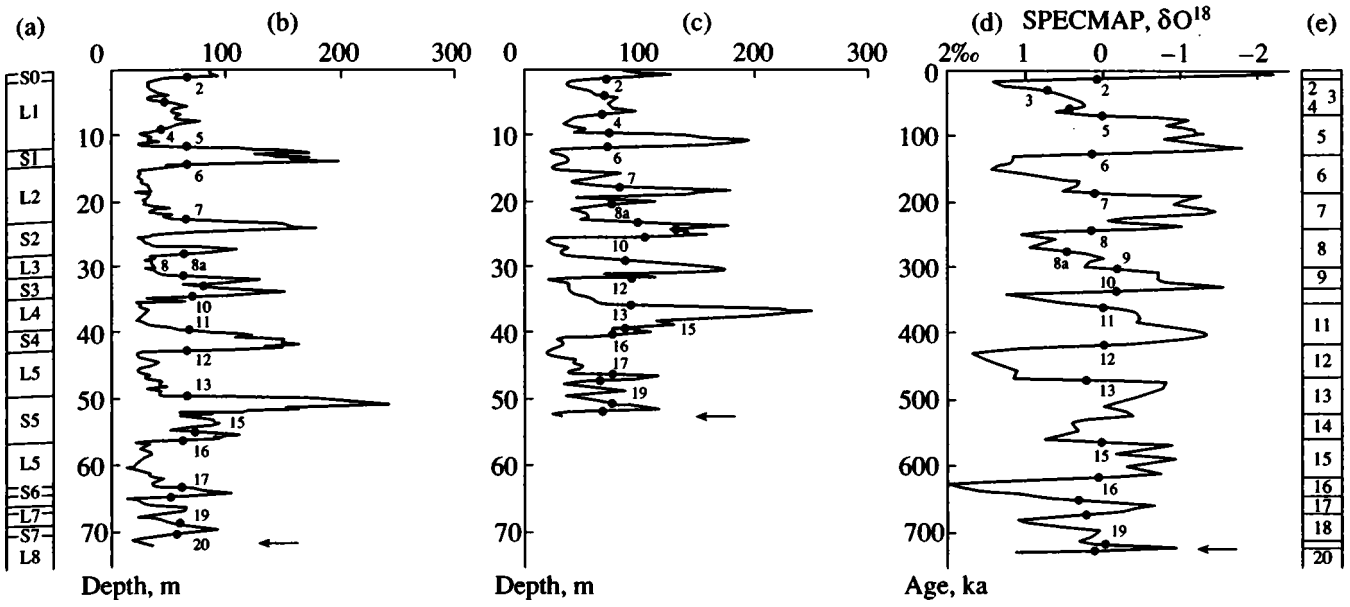

Fig. 1. Lithologic units (a) and magnetic susceptibility (2×10^{-5} SI units) variations in the Xifeng (b) and Luochuan (c) sections correlated with the SPECMAP oxygen-isotope scale (d) and OI stages (e): (L) loess, (S) soil. Large dots mark peaks of OI Stages and their time-equivalents in loess sections. Arrows indicate the position of the Matuyama–Brunhes reversal; all plots after Kukla *et al.* (1988).

variations was inadequately assessed, and interrelations between MS variations and paleoclimatic changes were improperly understood and only declared in works by Heller and Liu (1984, 1986). It is important, however, to substantiate the paleoclimatic nature of each significant minimum and maximum of κ values for the correctness of the spectral analysis.

Similar drawbacks are characteristic of work by Chlachula *et al.* (1998) who studied the Kurtak section in the Ob' River valley (Siberia) and revealed inverse MS parameters characterizing buried soils and loess beds, as κ values turned out to be higher in the latter (Fig. 2), like in the LF sections of Alaska (Beget and Hawkins, 1989; Beget *et al.*, 1990). Chlachula *et al.* (1998) connect κ variations in loess beds with climate-related changes in the wind intensity, giving no proofs in favor of this "Alaskan model" that does explain κ values in loess and soil. As was shown (Bol'shakov, 1997, 2000), the wind-induced MS variations have not been properly justified even by authors of the Alaskan model (Beget and Hawkins, 1989; Beget *et al.*, 1990). In particular, the MS increase in sediments of coarser grain size (Beget *et al.*, 1990) could result not only from eolian, but also from fluvial reworking of sedimentary material. Moreover, the model does not take into account the effect of chemical changes, which are established in loess beds of Alaska (Beget and Hawkins, 1989), on the κ value. That is why the correlation of κ values in the Kurtak section (Siberia) with those in the piston core V21-146 of Pacific sediments seems to be unconvincing. Similarly doubtful are substages distinguished (Fig. 2) in the OI Stage 5 (Chlachula *et al.*, 1998). Substages like 5a and 5b (Fig. 2c) can be distinguished as well in the overlying κ minimum, at the depth of about 10 m, the more so as stratigraphic position and age of BS horizons has not been substantiated in this section.

Thus, data of MS measurements in LF sections of China, Siberia, Alaska, and some other regions imply that *ratio* κ_{bs}/κ_l *characterizing sequences of buried soils and loess beds varies in space and can be equal, greater or less than one.*

In view of frequently exaggerated (for particular situations) paleoclimatic implications of MS measurements in loess and soil sections, of special interest are works, in which κ variations are directly correlated with paleoclimatic curves obtained on the basis of palynological or paleontological data. For instance, Bolikhovskaya *et al.* (1999) studied one of the characteristic LF sequences exposed in the Strelitsa reference section of the central Russian Plain. The correlation of paleoclimatic variations, inferred from palynological data, with data of MS measurements was a main objective of their work. Unfortunately, the interpretation and resulting inferences appeared to be controversial, as it is evident from the following principal conclusion: "In the Strelitsa section, like in other LF sequences, maximums of specific initial magnetic susceptibility (denoted as χ, V.B.) correspond to warm interglacial periods, whereas minimal or lower values of this parameter mark the cold glacial epochs. Paleoclimatic variations appear to be smoothed in initial magnetic susceptibility profiles lacking endothermal cooling phases. The role of gleying that distorts the paleoclimatic record in magnetic profiles is noticeable"

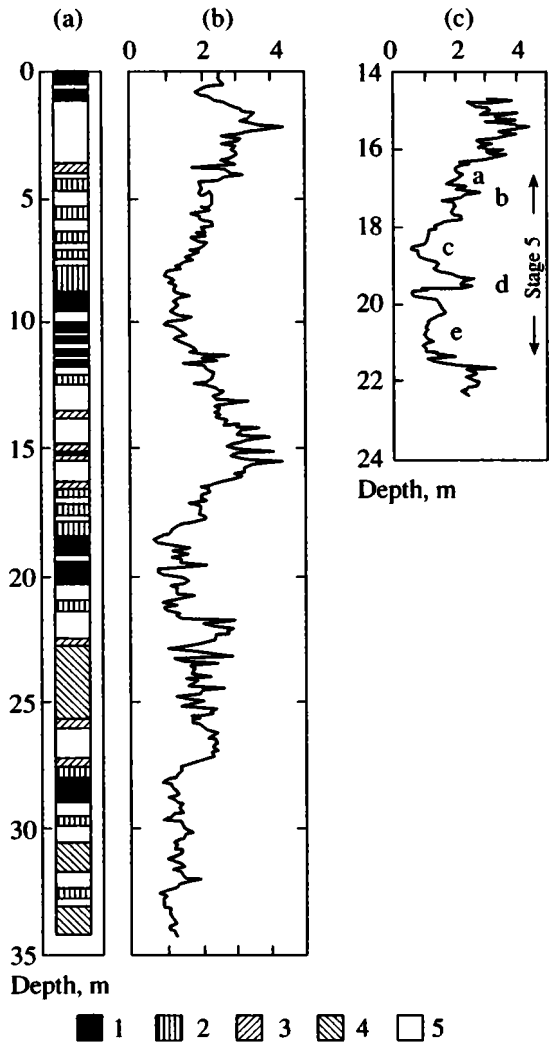

Fig. 2. Lithological column (a), magnetic susceptibility variations in the Kurtak section (b), and OI Stage 5 divided into substages 5a–5f (c) (after Chlachula *et al.*, 1998): (1) chernozem; (2) brown soil; (3) leached gley soil; (4) colluvial loess; (5) eolian loess (specific magnetic susceptibility in 10^{-6} m^3 kg^{-1}).

(Bolikhovskaya *et al.*, 1999, p. 690). It is evident that the initial inference (first sentence) is considerably weakened by conclusions following it. Moreover, the second and the third sentences are also controversial, conflicting with each other. The second one suggests that χ variations are paleoclimatic in origin, whereas the third sentence means that the paleoclimatic record in magnetic profiles is distorted by gleying. It would be logical to speak in this case about a distorted MS record of paleoclimatic events that actually implies defects of the record. My conclusion can be convincingly confirmed, probably for the first time, by concrete data presented in the work by Bolikhovskaya *et al.* (1999) (Fig. 3). For instance, transitional χ values correspond to soil horizons Romenka AC (interglacial phase) and Kamenka AI^1 (glaciation phase), and the average χ value of the AI^1 horizon is not lower than that of the AC horizon, as one can seen from the plotted χ variations. (I use the term transitional values instead of higher or lower values used in the original work, because the latter are misleading if the reference value is not indicated. For instance, values lower than a maximum may be easily greater than values specified as higher than a minimum.) Further, minimal χ values of the horizon B of the Inzhava soil correspond to the interglacial period, and transitional χ values characterizing the Korostel loess and AI^1 horizon of the Voron soil mark glaciation episode. Moreover, the χ maximum of AI horizon has no "lag behind the climatic optimum," as is stated in the work, but corresponds to the endothermal cooling phase of the Muchkap Interglacial, the beginning and climatic optimum of which are characterized by the minimal and transitional χ values (B and AB horizons of the Voron soil, Fig. 3). The ultimate refutation of the first sentence in above conclusion is a merit of experts in paleomagnetology (Virina *et al.*, 1998). They argued that ratio between χ values of loess and BS could be greater or less than one in different regions (I already mentioned this above) and pointed out that the χ value minimum is recorded in the podzolic horizon of the Mezin soil correlated with an interglacial stage. Another important fact is also presented in the work by Virina *et al.* (1998). Low χ values are typical of gley soils in the loess–soil sequence below morainal deposits that has not been considered in the work by Bolikhovskaya *et al.* (1999).

Thus, the analysis of two publications (Bolikhovskaya *et al.*, 1999; Virina *et al.*, 1998) reveals diverse relations between κ values of loess and soil horizons in the Strelitsa section and climatic events recorded here. In contrast to authors who insufficiently substantiated their controversial inferences with respect to paleoclimatic implications of κ values, I arrived at two following conclusions: (1) ratio κ_{bs}/κ_l can substantially vary in space and with time (might be greater of less than one); (2) variations of κ values in the LF sections *not always reflect* climatic changes recorded in the latter and could be controlled by nonclimatic factors.

Clearly, the second conclusion does not deny that the κ value *may reflect* paleoclimatic changes. For instance, this is evident from work by Pospelova and Levkovskaya (1994), in which concurrent variations of MS parameters and palynological data in a soil bed the Pogrebya section (Moldavia) are established.

The work by Maher *et al.* (1994) is another example of the comprehensive analysis used to establish κ variations in soils and corresponding climatic changes, and to apply the obtained data for a concrete paleoclimatic interpretation. In this work, the magnetic susceptibility of recent soils of the Loess Plateau (China) was correlated in a mathematical form with amount of precipitation. Both parameters increase southeastwards. Their quantitative connections were used to assess variations of precipitation in the past. As the authors themselves

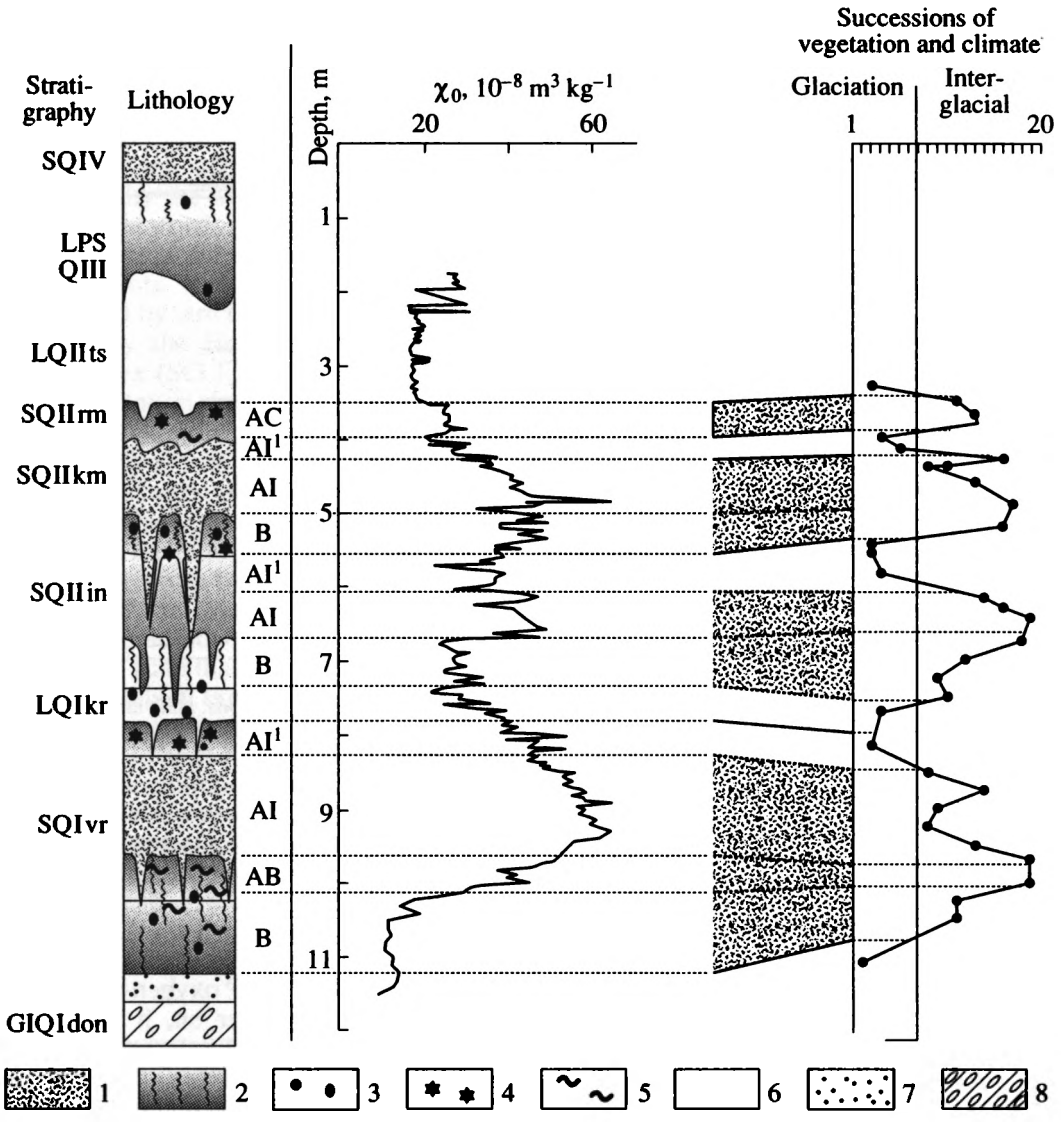

Fig. 3. Stratigraphy, lithology, κ variations, and paleoclimatic curve characterizing the Strelitsa section (after Bolikhovskaya *et al.*, 1999): (1) horizons A of buried soils; (2) horizons B of buried soils; (3) mole runs; (4) ferruginate patches; (5) gleying; (6) loess; (7) sand; (8) Don morainal bed. Letter symbols: (ts), Tyasma, (rm) Romenka, (km) Kamenka, (in) Inzhava, (kr) Korostel, and (vr) Voron loess and soil horizons. Shadowed areas near paleoclimatic curve correspond to interglacial stages.

note, the precipitation value they estimated for the last interglacial stage substantially exceed those obtained on the basis of other methods, in particular, by using the model of the general circulation in the atmosphere. In my opinion, this is a consequence of two essential drawbacks of the used approach (Maher *et al.*, 1994). First, assuming that κ values in soils are controlled by pedogenesis, they admit the soil formation to be dependent on the amount of precipitation only. It is known, however, that pedogenesis depends, for instance, on the sedimentation rate. Under a higher rate of eolian material sedimentation, the soil formation will be suppressed, and the κ value in soils will be lower. This inference agrees well with data on the Loess Plateau of China: κ values increase, as thickness of loess–soil horizons decrease here in the direction from Lanzhou to Luochuan (Burbank and Li, 1985; Kukla *et al.*, 1988; Hus and Han, 1992; Evans and Heller, 1994). Hence, formula describing dependence of κ value in soils on the amount of precipitation should include parameter of sedimentation rate as well. Accordingly, the κ value dependence on the amount of atmospheric precipitation remains ambiguous, since we do not know what kind of relations, if any, connects sedimentation rate and atmospheric precipitation, and how constant they are. One more complexity is a possible effect of temperature variations within the Loess Plateau, which likely take place and may influence pedogenesis as well, and, hence, the κ value of soils.

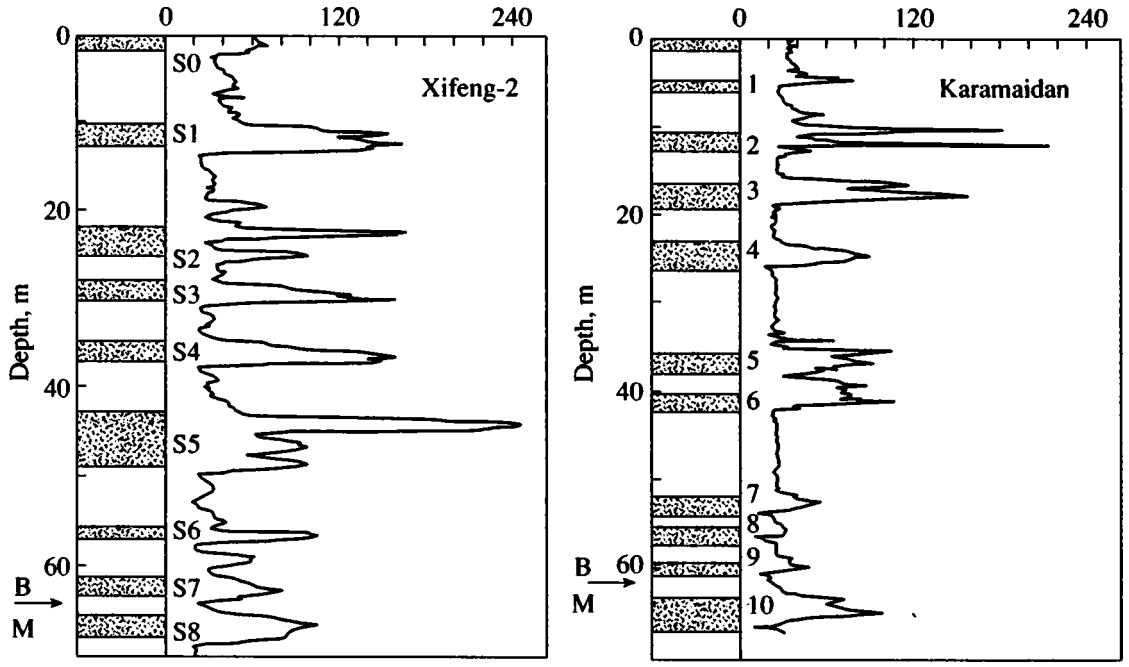

Fig. 4. Magnetic susceptibility (10^{-8} m^3 kg^{-1}) variations in the Xifeng-2 and Karamaidan sections; soils enumerated from the top downward are shown in black, and arrows show position of the Matuyama–Brunhes reversal (after Shackleton *et al.*, 1995, Fig. 1).

Second, Maher *et al.* (1994) ignored possible transformations of magnetic minerals in soils after their burial. I reported on such (at least, qualitative) transformations in buried soils of Bulgaria, similar in magnetic characteristics to soils of China (Bol'shakov, 1996). The possibility of secular changes in BS magnetic properties is implied by obvious diagenetic effects resulting, for instance, by the post-burial decrease of humus content in BS that is progressing with time (Kriger, 1965). Such a process, if it takes place, eliminates possibility of paleoclimatic interpretations based on the comparative analysis of κ value variations in buried and recent soils.

Kukla *et al.* (1988) initiated consideration of MS parameters as chronostratigraphic indicators. Assuming that magnetite concentration, i.e., the κ value in loess or soil, depend on the loess accumulation rate, they proposed a model, according to which the natural magnetite dust from the high atmospheric layers accumulated with a *constant* rate on the Earth's surface. Consequently, if the κ value and thickness of sedimentary layer are known, one may determine the time interval of sedimentation, provided the continuous character of sedimentation. Another necessary condition is presence of two dated levels in the section (usually the beginning of the Holocene and the M/B reversal), and age of layer m can be determined then by formula:

$$T_m = T_1 + \left(\sum_{i=1}^{m} a_i\chi_i\right)(T_2 - T_1)\left(\sum_{i=1}^{n} a_i\chi_i\right)^{-1}, \quad (1)$$

where T_2 and T_1 are maximal and minimal ages of levels dated in the section; n is amount of layers between dated levels; a_i and χ_i are thickness and magnetic susceptibility of i-layer. Using this model, Kukla *et al.* chronologically correlated the Luochuan and Xifeng sections of the Loess Plateau in China. Their correlation with the SPECMAP oxygen-isotope scale was also adequate (Fig. 1). Later on, it was shown, however (Bol'shakov, 2000, and references therein), that the model under consideration is invalid because of κ variations in response to conditions of pedogenesis. Nevertheless, Shackleton *et al.* (1995, p. 2) attempted to revive the considered model admitting the following mechanism of κ variations: "To a first order approximation, the flux (formation and/or deposition) of the susceptibility carrier is considered constant...". It is evident that the assumed constant "formation" of susceptibility carrier is as much unsubstantiated as the previous condition of constant subaerial deposition. In the last work, chronological correlation between the Xifeng-2 section of China and the Karamaidan section of Tajikistan was based on the MS measurements (Fig. 4). Figure 4 well illustrates the impossibility of correlating these two sections located far apart by means of κ variations as such, although this is admissible for the Loess Plateau section in China (Fig. 1).

The above discussion means that, despite the originality, the MS time scale elaborated by Kukla and his colleagues is invalid, since it is based on incorrect physical prerequisites. According to postulated mechanism, the κ value depends only on sedimentation rate: the lower the rate (in soils) the greater the κ value. Accord-

ing to given comments, this mechanism may work only in regions with well-drained soils (for instance, in China, but not in Alaska or Siberia). Nevertheless, the mechanism is inappropriate even for China, because κ values depend, according to data mentioned above, on one more parameter—on precipitation value (Maher *et al.*, 1994). Then why the MS time scale yields reasonable results (Kukla *et al.*, 1988, 1990; Shackleton *et al.*, 1995)? Here, we should take into consideration two important moments. First, all the age values are originally constrained by strict time boundaries: by the M/B reversal and by the Holocene base or by the Mikulino soil complex (SC 1 in Fig. 4). Provided the absence of sedimentation breaks, these limitations ensure a reasonable dating of loess and soil horizons within the corresponding time intervals. The dates will be even more precise if variability of the sedimentation rate can be taken into account. This is just the second moment conditioning the reasonable dating by formula (1), which puts the κ value into correspondence with the formation time of the given sequence. The correspondence is solely qualitative, as I should repeat that we have no reason to state that soils with low κ values (for instance, lower soils in Figs. 1 and 4) accumulated faster than soils with greater values of this parameter. Similarly founded, or to be more exact, unfounded, one may suppose that sedimentation rate by accumulation of loess horizons is greater by factor of two, three, four, etc., than soil formation. In this case, the age can be calculated from the next formula:

$$T_m = T_1 + \left(\sum_{i=1}^{m} a_i k_i\right)(T_2 - T_1)\left(\sum_{i=1}^{n} a_i k_i\right)^{-1}, \qquad (2)$$

where k_i equals 1 for even-numbered i, or 2, 3 and 4 for odd-numbered i, if we begin to count from the loess layer (accordingly, loess corresponds to odd-numbered I). It is also possible to specify various k_i values for soils with different profiles. Then we can precise again ages of horizons as compared to the alternative approach implying the constant sedimentation rate. In fact, dating by formula (2) is more general than by formula (1) and can be applied to both the automorphic and hydromorphic soils.

Reverting to work by Shackleton *et al.* (1995), I should point out that the MS time scale yielded poor correlation of the Xifeng-2 and Karamaidan sections with the OI scale. Therefore, they had to adjust the MS variation profile to the OI scale. The unavoidable consequence of adjustment, which is not discussed in above publication, is the necessity to displace the M/B reversal from the loess horizon 7 to overlying soil horizon S 7 (Fig. 4).

Closing the discussion of results obtained by Shackleton *et al.* (1995), let us consider their main inference. It implies that of eolian processes in China and Tajikistan depended on dynamics of the global ice volume changes, which is recorded by OI variations in deep-sea sediments, to a greater extent than on theoretical variations of insolation. This inference is based not on interpretation of measured MS parameters, but on adjustment of climatic cycles recorded in the sections to the OI scale. The inference seems reasonable from theoretical side as well. The point is that, as shown in (Bol'shakov V. and Bol'shakov P., 1999), Milankovitch theory has some essential drawbacks, and insolation curves calculated on the basis of this theory disagree with OI data and cannot be used for paleoclimatic interpretations. The last work *directly* demonstrated a decisive effect of insolation, controlled by variations of the Earth's orbital elements, on the global ice volume changes during the Pleistocene. These changes determined, in turn, alternation of glacial and interglacial periods detectable in OI records data, and accompanying reorganizations in the system of atmospheric and oceanic circulation. During glacial periods, for instance, the atmospheric circulation should be intensified because of the greater latitudinal gradient of temperature. That is why the loess formation indicative of the intense atmospheric circulation is directly correlated with glaciations (if we leave aside changes in humidity of a given region).

DISCUSSION

Some works, in which the MS measurements are used to reveal paleoclimatic conditions during loess and soil formation, are considered above. It is possible to state *a priori* that global climatic changes, which influence sedimentation conditions, should be reflected in variations of magnetic properties of LF sequences. These variations are also characteristic of recent soils. For instance, Babanin (1971) established that soil horizons A and C have variable magnetic susceptibility: ratio κ_A/κ_C varies from one or lesser value in sod–podzolic soils to three–four in chernozem and chestnut soil. However, not only climatic but many other factors independent of climate control the κ value of soils. I should again mention the works by Bolikhovskaya *et al.* (1999) and Virina *et al.* (1998), according to which gleying may substantially decrease the κ value in soils, thus reducing to zero the global climatic signal that could be recorded otherwise. On the other hand, global climatic changes themselves are very diverse and complex, variably affecting the environmental parameters, such as temperature, humidity, and others. These changes are nonuniform in different regions of the planet. They all are responsible for the space-and-time variations of κ values in loess and soil.

Hence, the effect of paleoclimate on magnetic properties of sediments cannot be refuted. It is difficult to predict, however, the quantitative changes of magnetic susceptibility in particular LF sections in response to climatic factors, if we do not know the concrete relationships between κ value variations and changes in environmental conditions. It is even more difficult to solve the reverse problem, i.e. to reconstruct paleocli-

matic changes based on MS variations. It is pertinent to ask here what the MS measurements mean in terms of paleoclimatic reconstructions, which are under discussion in many works? The magnetic susceptibility parameter represents a particular value and nothing else. Hence, *the application of MS parameters for paleoclimatic reconstructions is a procedure independent of space and time that means the direct correlation of particular* κ *value intervals with corresponding climatic parameters* (temperature, humidity, direction and intensity of wind, which affect sedimentation rate). Works considered above show that this problem cannot be solved at the global scale, since κ values of loess and soil, as well as ratios κ_{bs}/κ_l depend on local and regional climatic factors. Could it be solved within a separate region, in which the climate-induced κ variations are uniform, as for instance, in sections of the Loess Plateau of China? It is a common practice to consider paleoclimatic changes in generalized qualitative terms like glaciation or interglacial stage, but the problem I pose now is much more complicated. To solve it, it is necessary to establish first the concrete dependence of κ values on every significant climatic factor. Even in the case of a considerable progress in this direction, there is no way to ensure the unambiguous correlation between climatic changes and MS variations and, hence, the unambiguous solution of the reverse problem. The κ value depends, at least, on three parameters, and the unequivocal solution is possible only if these climatic parameters (temperature, moisture, wind intensity) are interrelated in a definite manner. The question whether or not such interrelation is definite and invariant with time is not useless. In the work by Banerjee *et al.* (1993), it is assumed that variations in magnetic properties of Holocene soils, in contrast to soils of earlier interglacial stages, are caused by concurrent changes in the paleomonsoon circulation. Clearly, this should change the regional humidity and eolian transport of sedimentary material in the region, and, hence, the new variations of κ values.

Thus, we are far away from the moment, when correlation between κ values and global climatic changes determining specific dependences like κ–temperature or κ–humidity could be established for particular sections, and when the concrete physical mechanisms enabling the correct interrelation could be understood (if it is possible to found them). Only after positive results obtained on this way of comprehensive study, we may pass from speculations to the correct interpretation of MS measurements in paleoclimatic terms. Only such a strict scientific approach may facilitate solution of the discussed reverse problem, and only it may create a basis for the correct determination of paleotemperatures or paleohumidity with the help of MS measurements, in contrast to immature solutions presented in some works (Maher *et al.*, 1994; Heller, 1998). One should remember therewith the possible diagenetic transformations of magnetic minerals after their burial in loess and soil horizons. If this and another non-climatic factors have a decisive effect on magnetic susceptibility, then the correct solution of the reverse problem appears to be impossible.

Taking into account all complications, which appear when we try to correlate the continental LF (MS variation curve) and marine (OI curve) sedimentary sequences, I advise to use instead either the reference loess (glacial) and soil (interglacial) horizons, or traditional paleoclimatic curves (arid and cold versus humid and warm intervals). Correlations of this kind were carried out in some works (Liu *et al.*, 1984; Ding *et al.*, 1991; Dodonov and Baiguzina, 1995), and the results obtained are as good as those based on interpretation of MS measurements. For instance (Heller and Liu, 1984, 1986; Kukla *et al.*, 1988; Liu *et al.*, 1992), the number of κ-stages distinguished within the Brunhes Chron and below the Olduvai Subchron of the Loess Plateau sections is variable. At the same time, the number of OI Stages correlative with loess and soil horizons below the Olduvai Subchron (Ding *et al.*, 1991) is closer to that presently accepted in contrast to correlations based on MS measurements. The imperfect correlation between deep-sea and continental sedimentary sequences is not only a consequence of mistakenly distinguished stages based on MS measurements. Some inconsistencies can be associated with incompleteness of continental geological records deposits and with ambiguous subdivision of soil complexes (SC). The comparative analysis of works (Dodonov, 1986; Dodonov and Baiguzina, 1995; Shackleton *et al.*, 1995; Liu *et al.*, 1984; Kukla *et al.*, 1988) suggests that the distinguishing of SC ranks (interglacial, interstadial) and individual soil horizons is a subjective procedure, on which the number of MS stages or soil complexes used in correlation depends.

One more problem is positioning of the M/B reversal in LF sequences. The OI scale and paleomagnetic data on deep-sea sediments are very helpful in this case, but the last word should be that of experts in paleomagnetism and magnetism of rocks.

CONCLUSION

Works analyzed earlier (Bol'shakov, 2000) and above lead to the conclusion that interrelations between MS variations in LF sequences and global climatic changes are ambiguous. Climatic impacts can (or cannot) be manifested in different manner not only in LF sequences of various regions, but also in different parts of one section. Two main reasons responsible for this are as follows: (a) MS variations in rocks sequences are controlled by various, frequently independent factors, some of which have no direct relation to global climatic changes; (b) global climatic cycles of the Pleistocene dissimilarly manifested themselves in space and with time.

Summing up the available attempts to use the MS parameters of LF sequences for paleoclimatic recon-

structions, I should point out the following. The MS variations are ineffective for establishing qualitative climatic changes of the "glaciation–interglacial stage" type, since a more simple indicator of these events is color of corresponding horizons stable in space and with time. The light color of loess usually indicates a glaciation, whereas the dark color of soil points to an interglacial stage. Attempts to use quantitative MS parameters as indicators of paleoclimatic changes (in temperature, humidity, and duration of climatic cycles), have no proper methodological substantiation so far, since we do not know in detail how the κ value depends on all factors controlling sedimentation conditions. Even in the case of automorphic soils of China, it is difficult to interpret the MS measurements, because it is still unknown what mineral, magnetite or maghemite, is responsible for the magnetic susceptibility increase in buried soils (*Pedogenesis and paleoclimate...*, 1994).

Hence, it is clear that the idea of special importance of MS measurements in LF sequences that is declared by some researchers, and has turned in their minds into a material power forcing them to make unjustified statements, is not a real knowledge as yet. Attempts to use the MS measurements even in those cases, when they were unnecessary, imply that we have no adequate method to study the Pleistocene LF formations.

Application of MS measurements for geological correlation of different sections can be successful in regions with rather uniform sedimentation conditions. This concerns not only LF sequences but other sedimentary rocks as well (Thompson *et al.*, 1980; Verosub and Roberts, 1995). The deviation of κ values from those typical of the given region can be regarded as an express-signal informing us that regional sedimentation conditions changed in the studied locality. Such an express-signal, if available, implies that we have to give special attention to horizons "anomalous" in terms of the MS parameters, and that they deserve a comprehensive study (application of new research methods or more detailed sampling). The κ value is a very sensitive indicator characterizing the lithological peculiarities of deposits, which are often invisible. That is why the MS measurements can be successfully used for revealing lithological distinctions of sedimentary rocks. In particular, the MS method has been widely applied to study the LS horizons and sedimentary core sections from sea bottom. For instance, it helps to distinguish horizons of poorly developed buried soils during the field work.

Thus, only the methodologically correct scientific approach to investigation of magnetic susceptibility in sedimentary rocks can turn the MS method into a useful tool complementing traditional methods of paleogeographic studies.

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The manuscript should be attentively corrected and signed by author(s). Clear and easily understandable writing style is required for an adequate translation into English. A letter of transmittal from host organization, full author(s) name(s), and possibilities of communication (phone numbers, post and e-mail addresses) should be attached. Submission of text in printed and electronic forms is required.

Body text of the manuscript (two copies without handwritten corrections) should be printed double-spaced on a white paper, size A4 (297 × 210 mm), with the left margin 25 mm wide or more and 29 to 30 lines per a page. All pages should be numbered in order (top center of each page). Figures, tables, and photographs submitted separately should not exceed the A4 size.

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Member of the Russian Academy of Sciences, Doctor of Geology and Mineralogy

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